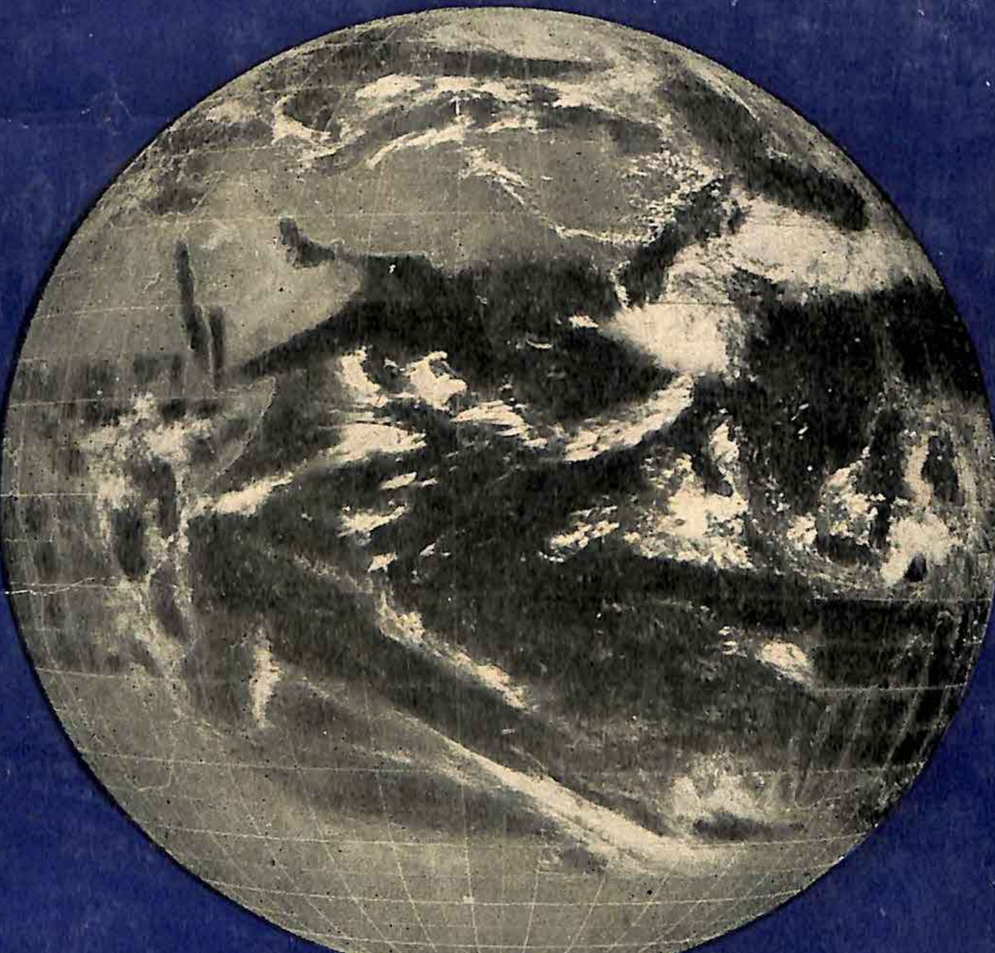
India—The Land and the People

THE MONSOONS

P.K. Das



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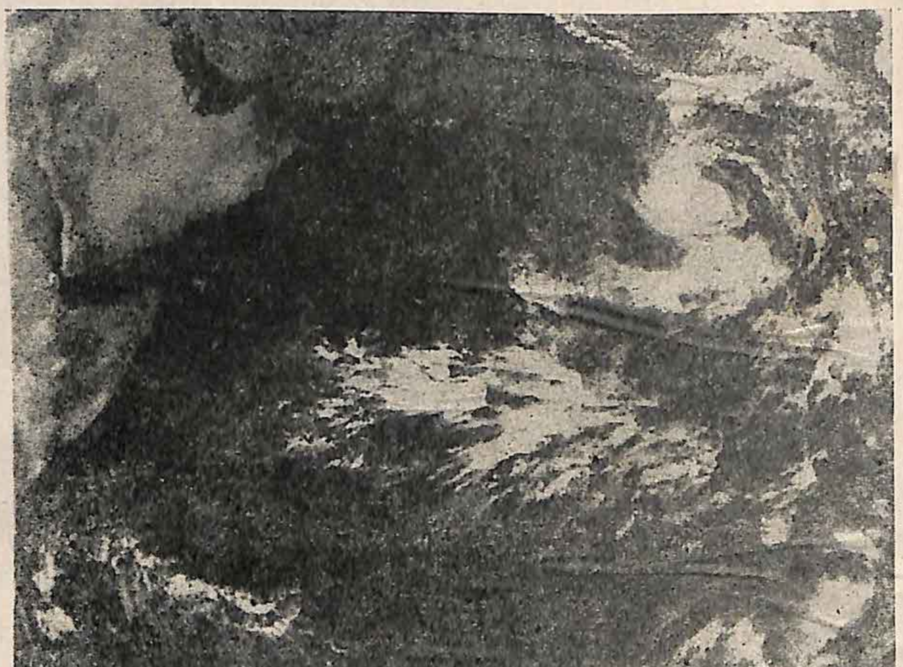

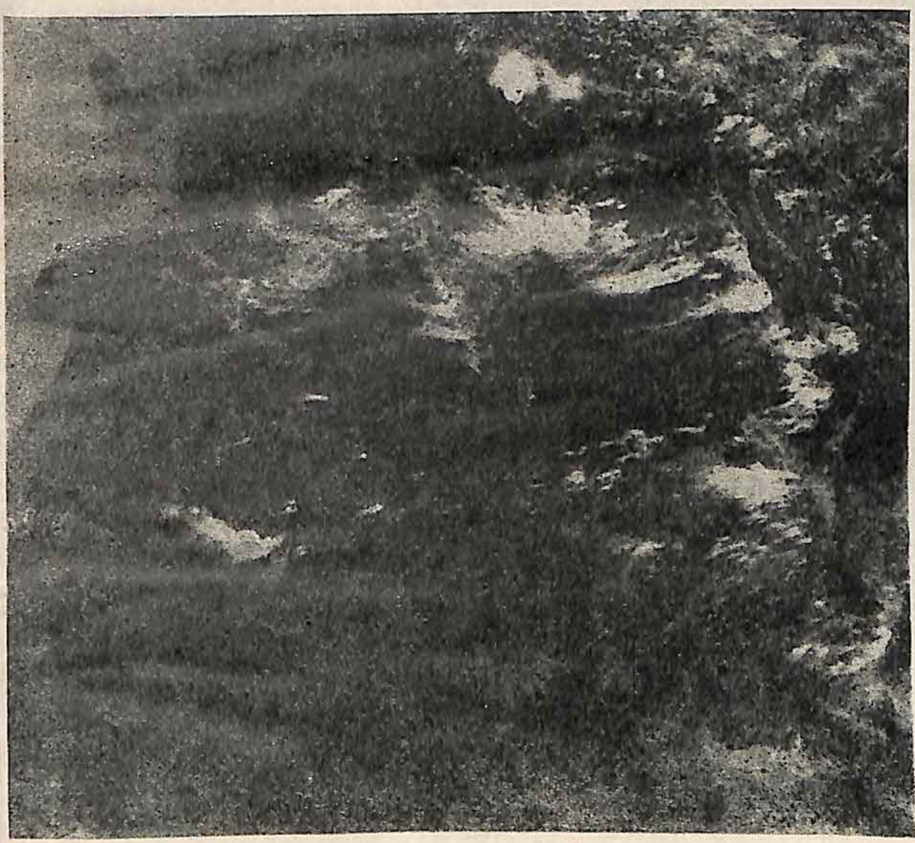

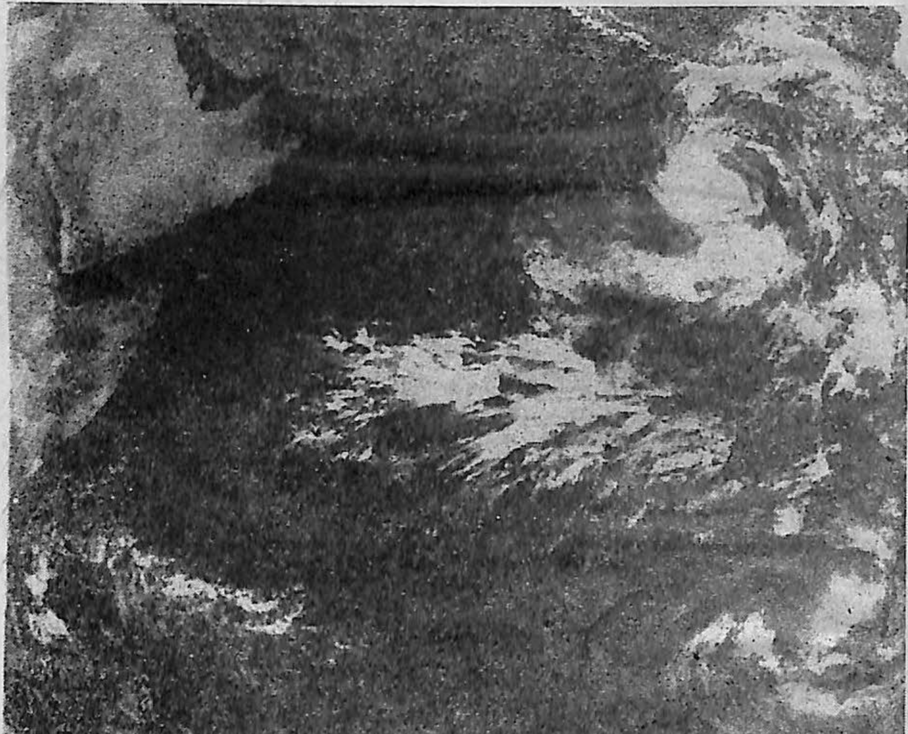

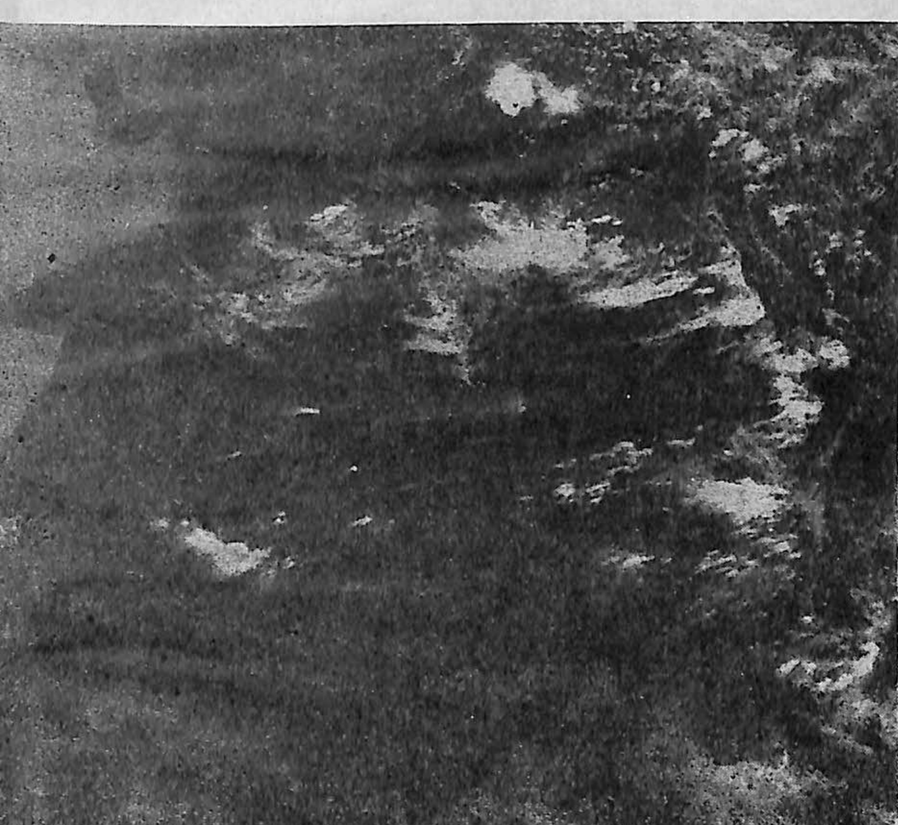

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P. K. Das

NATIONAL BOOK TRUST, INDIA

First Edition 1968 (*Saka* 1890)
Second Edition (Revised) 1988 (*Saka* 1910)

Rs. 18.00

Published by the Director, National Book Trust, India,
A-5, Green Park, New Delhi-110016

PREFACE

THE GENEROUS RESPONSE TO the first edition of the book has encouraged me to bring out a second edition. As with the first edition, this is an introductory text on Monsoons meant for interested students and readers who would like to learn more about the Indian summer monsoon.

Eighteen years have passed after the first appearance of the book. Interest on the physics of global monsoons has considerably increased during this period. This is largely an outcome of several international expeditions designed to study monsoonal circulations. The last one of the series, the Monsoon Experiment (MONEX) of 1978-79, provided new data and observing platforms. The recent geostationary satellite launched by India and the Indian expeditions to the Antarctic have added to the resurgence of interest on Monsoons. These developments have prompted me to rewrite many parts of the earlier edition. I should like to state that much of the Monex data still await detailed analysis ; consequently, many inferences are tentative at this stage. I have endeavoured to interpret observations as I saw them, and to indicate areas where our understanding is not yet clear. The exposition is in simple language which should make it intelligible to as wide an audience as possible.

I wish to acknowledge the courtesy and generosity of the National Book Trust, India, especially Mr. Arvind Kumar, the Director for his patience and understanding.

Many of my colleagues rendered valuable assistance by giving me the benefit of their experience. It is possible that some will have a different view from mine but, notwithstanding divergence

of opinion, I wish to acknowledge their help.

Several references have been made in the text to classical papers at end of the book for each chapter.

Mr. M. G. Gupta and Mr. S. D. Gaur spared no pains in preparing and typing the manuscript. Without their assistance it would have been difficult to bring out this edition. Their help is gratefully acknowledged.

New Delhi.
March 30, 1986

P. K. Das

CONTENTS

LIST OF DIAGRAMS

To

My Father

Who asked me to write it;

My Mother

Who will see it no more,

And Mr. Jack Gibson

Who taught me to observe

The atmosphere;

This book is dedicated.

CHAPTER 1

INTRODUCTION

Historical background

THE WORD MONSOON OWES its origin to an Arabic word meaning 'season'. It was used by seamen, several centuries ago, to describe a system of alternating winds over the Arabian Sea. These winds appear to blow from the northeast for about six months and from the southwest for another six months.

One of the oldest literary works of the Aryans is the Rg-Veda. It contains over a thousand hymns and songs sung in adoration of the deities. There are many references in this work to the rivers, the mountains, the sea and the desert in the vicinity of north-west India, where the earliest Aryans settled on coming to India. Scholars of history have placed the Rg-Veda between 1200 and 500 BC ; but some place it between 4000 and 6000 BC.

Although the word *Monsoon* was coined much later, M. V. Unakar, an Indian scholar, drew our attention nearly fifty years ago to many verses in the Rg-Veda which praise Parjanya, the God of rain and the creator of plants and living creatures. The following verse, reproduced from Mr. Unakar's treatise, provides a description of what could be the monsoon rains in the north-west India :

Thou hast poured down the rain-flood, now withhold it ;
Thou hast made desert places fit for travel,
Thou hast made herbs to grow for our enjoyment,
Yes ! thou hast won praise from all living creatures.
They who lay quiet for a year, the frogs,
Have lifted up their voice, the voice which Parjanya inspired ;
Soon the rain time in the year returneth.

The last few lines are interesting because they refer to the return of a "rain-time", that is, the rainy season of the year. The fact that there was a period when it rained heavily every year was observed by the early Aryan settlers.

In later periods there occur descriptions of the monsoon in the early history of countries which had a maritime interest over the Arabian Sea and the neighbouring parts of the Indian Ocean. There occurs a description of the monsoon in the *Periplus of the Erythraean Sea* written about 60 AD by an unknown Greek sailor. A *periplus* is a navigator's pilot-book, and *Erythraean Sea* was the name given by the ancient Greeks to the Red Sea, the Persian Gulf and the Arabian Sea.

Around 400 AD Fa Hsien, a Buddhist scholar visited India from China. Shortly after his visit he wrote a "Record of the Buddhist Kingdoms", in which there was a reference to the winter monsoon encountered during a voyage along the east coast of India. Perhaps the most beautiful description of monsoon clouds appear in a Sanskrit classic, *Meghdoot*, by Kalidasa around the fourth century. In this book, Kalidasa described the arrival of the monsoon over Ujjain (in Madhya Pradesh) on the first day of Asadha (15th of June); this is a surprisingly accurate estimate of the date on which monsoon rains arrive over this part of India.

Near about the tenth century, Al Masu'di, an Arab geographer from Baghdad, wrote a book entitled the *Meadows of gold and mines of gems*, in which there was an account of the reversal of ocean currents over the north Indian Ocean. The book did not deal with "meadows" of gold or the "mines" of gems; but the title was coined by the author to "excite a desire and a curiosity after its contents and to make the mind eager to become acquainted with history".

Such historical anecdotes suggest how much the monsoon rains influenced the lives of men for nearly two thousand years over our sub-continent.

Monsoonal changes

While there is lack of agreement on a precise definition of the monsoon, the term is used to connote a seasonal wind, which blows with consistency and regularity during a part of the year, and which is absent or blows from another direction for the rest of the

year. Such seasonal changes of wind are primarily the result of differences in the quantity of heat received from the sun by different parts of the earth.

There is a striking difference in the response of the continents and the oceans to seasonal changes in solar energy. As a consequence of its chemical composition and the structure of soil, the conduction of heat into the earth is a slow process. In summer, for instance, only a shallow layer of a few centimetres of soil are heated by the energy received at the ground. Most of the solar energy received by the land is used up in heating the air rather than the earth's surface. On the other hand, solar energy is able to penetrate to much greater depths in the oceans because of the stirring which goes on under the action of the wind. The incoming solar heat penetrates to a depth of at least 200 metres because of turbulent transfer of heat. Consequently, a smaller part of solar heat is available for heating the air. The overall result is that the rise in temperature in summer is much less over the oceans than over the continents. The mean summer temperatures over land often exceed those over oceans along the same latitudes by as much as 5 to 10°C. In winter the situation is reversed, and the larger heat storage of the oceans leads to higher temperatures there than over land.

To raise the temperature of a unit volume of water requires much greater heat than is needed to warm an equal volume of air, and the land responds much faster to changes in solar heat input than the oceans.

In its generic sense then, we may describe the monsoon as a system of winds with the following features :

(*i*) A system, with marked seasonal wind shifts, caused by the differential heating of the land and sea, that is, by the different response of the land and ocean to incoming radiation from the sun.

(*ii*) Winds that are largely confined to the tropics, by which we mean the region between 20°N and 20°S on both sides of the equator.

(*iii*) Summer monsoons over the northern hemisphere may be thought of as the southeast trades or trade winds of the southern hemisphere which, on crossing the equator, are deflected to the

right by the earth's rotation and, as a consequence, approach the land areas from a southwesterly direction.

The question might be asked : If it was only a difference in the response of the land and oceans to incoming solar energy, why are monsoons not observed wherever large areas of land are surrounded by oceans ? The answer lies in the energy that we need to set up a reversal of winds on a scale comparable to the monsoons. Clearly, the temperature contrast between land and sea must be very pronounced and persistent, to be able to generate a monsoon. But, if the thermal contrast was on a smaller scale and lasted a short while, we still observe a phenomenon that is similar to the monsoon. This is a sea-breeze, which is a common phenomenon near a coast.

During the day, when solar insolation is large, the land becomes warmer than the sea. This generates a wind blowing from the sea to the land. During night, when there is no solar insolation, the land cools more rapidly than the sea. This results in a breeze from the land to the sea. The intensity of the sea breeze, or the reverse land breeze, depends on the thermal gradient between the land and the sea. The monsoons occur over the tropics, because it is this region of the world that receives most solar energy and is able to build up intense thermal gradients between land and oceans.

The similarity between monsoons and the land or sea breeze was noticed in a pioneering work by the English astronomer Edmund Halley in 1686. His remarkable work was later extended by George Hadley in 1735. He provided an explanation of the trade winds, which is essentially true even today. The circulation over the tropics is now named as a *Hadley Cell* in honour of George Hadley. These early landmarks in the study of monsoon winds were published in the Philosophical Transactions of the Royal Society of England.

C.S. Ramage, in a book entitled *Monsoon Meteorology*, suggests four features of monsoon winds :

(*i*) The prevailing wind direction should shift by at least 120° between January and July.

(*ii*) The average frequency of prevailing wind directions in January and July should exceed 40%.

(*iii*) The mean resultant winds in at least one of the months should exceed $3ms^{-1}$.

(*iv*) There should be fewer than one cyclone—anticyclone alteration every two years, in either month, over a five degree latitude-longitude grid.

These characteristics emphasise the seasonal nature of the change in wind direction, and the persistence of the wind regime in each season. The last definition, for example, ensures that the changes reflect the replacement of a persistent circulation by a reverse and equally persistent one. The change must not be a temporary one brought about by variations in the track of moving pressure systems, such as, tropical cyclones.

A precise definition of the monsoon is a matter of choice, because one need hardly frame a set of rules which embraces all facets of the monsoon. Ramage's definition, for example, considers only surface winds. By his definition, the deserts of Sahara should be a part of the monsoon regime of Africa, even if the rainfall there was quite different from northeast India—another monsoon regime. The former hardly receives any rain, but the latter has one of the heaviest rainfall of the world. Opinions differ on whether the monsoon should be defined by its rain-generating capacity, or by changes in global winds.

It is of interest to recall that the monsoon winds are most pronounced in the summer season of either hemisphere, that is, during the months of June, July and August in the northern hemisphere and in January and February in the southern hemisphere. During June, the trade winds from the southern hemisphere penetrate deep into the northern hemisphere towards India, the wide stretches of southeast Asia and to a lesser extent towards Africa. On the other hand, in January, the northeast trades move southwards into south America, east Africa and north Australia. During this period, a branch of the northeast trades sweeps across the south Bay of Bengal and generates precipitation over the southern half of the Indian peninsula. This is well-known as the winter or northeast monsoon of India. In very general terms, those areas of the earth, which experience large-scale air movements from a colder to a warmer hemisphere, are the principal monsoon lands of the world.

The monsoon currents associated with other continental masses of the earth are not so well marked as the Indian monsoon, but seasonal changes in wind direction are known to occur over north Australia, western and eastern Africa and southern U.S.A. A tendency for northwest winds to blow from the Atlantic into Europe in the summer months of June and July is sometimes referred to as an European monsoon. The monsoon is one of the main features of the Sahel region of Africa, because it provides the region with water from the atmosphere. During this part of the northern summer, the quantity of monsoon rain plays a decisive role in the struggle for survival of life and vegetation. As with India, the monsoon is important for the economy of many parts of western and central Africa. The mountains of the eastern coast of Africa are also important because they help to deflect the south-east trades towards India. As we shall see subsequently, the seasons of 'long' and 'short' rains over east Africa, which correspond to the summer and winter monsoons of the northern hemisphere, are generated not so much by differential warming but by seasonal shifts of trade winds.

For the present, our study will be mainly confined to the Indian summer monsoon, but we will discuss, briefly, the large-scale aspects of other monsoons.

The terminology that is used to delineate a zone of separation between the northern and southern hemisphere trade winds has had a chequered history. Many years ago the name "Inter-tropical Front" (ITF) was used to indicate a narrow zone in which air from the summer and winter hemispheres move towards the equator ; but this led to confusion because meteorologists in the middle latitudes were accustomed to using the word "front" to separate air-masses in fast moving low-pressure systems. To avoid the use of a "front", the term "Inter Tropical Convergence Zone" (ITCZ) is now widely used, but over western Africa the ITCZ is often replaced by an "Inter Tropical Discontinuity" (ITD) because the weather that is associated with the ITCZ over east Africa is different from what is observed over the western coast of Africa.

When we consider broad streams of air from both hemispheres moving equatorwards, we expect from considerations of continuity that the ITCZ should be a region of ascending air and

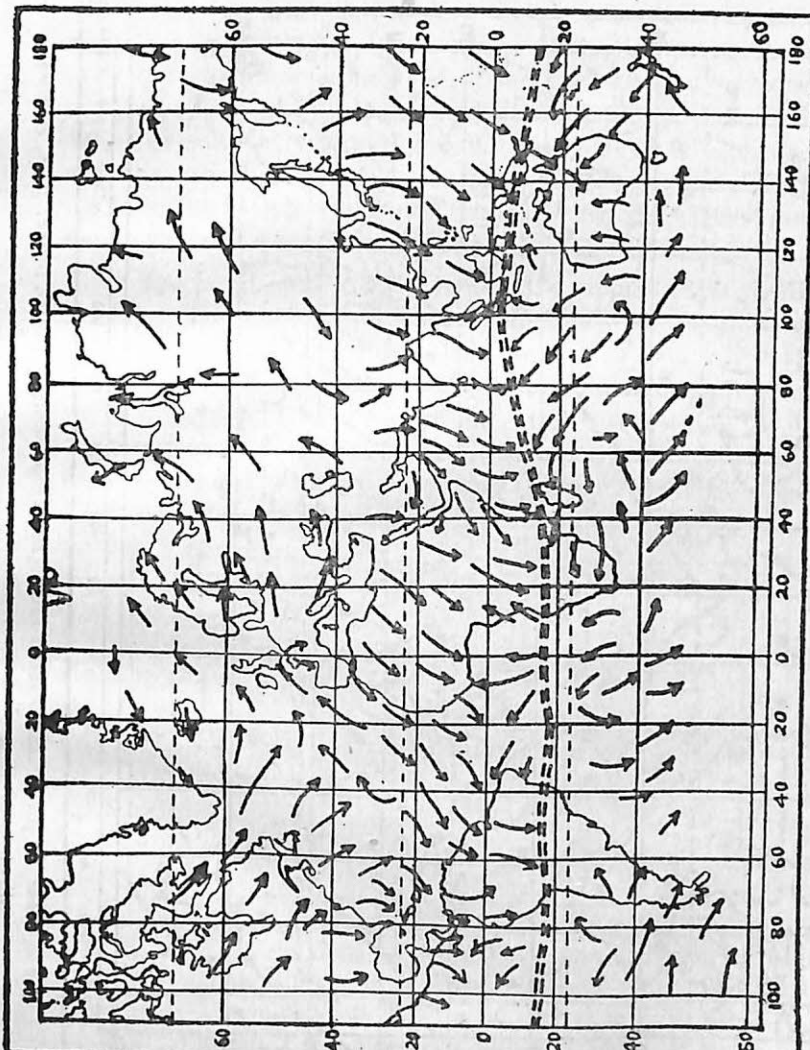

Fig. 1.1—Inter Tropical Convergence Zone (January) represented by thick broken lines. Arrows indicate prevailing wind directions.

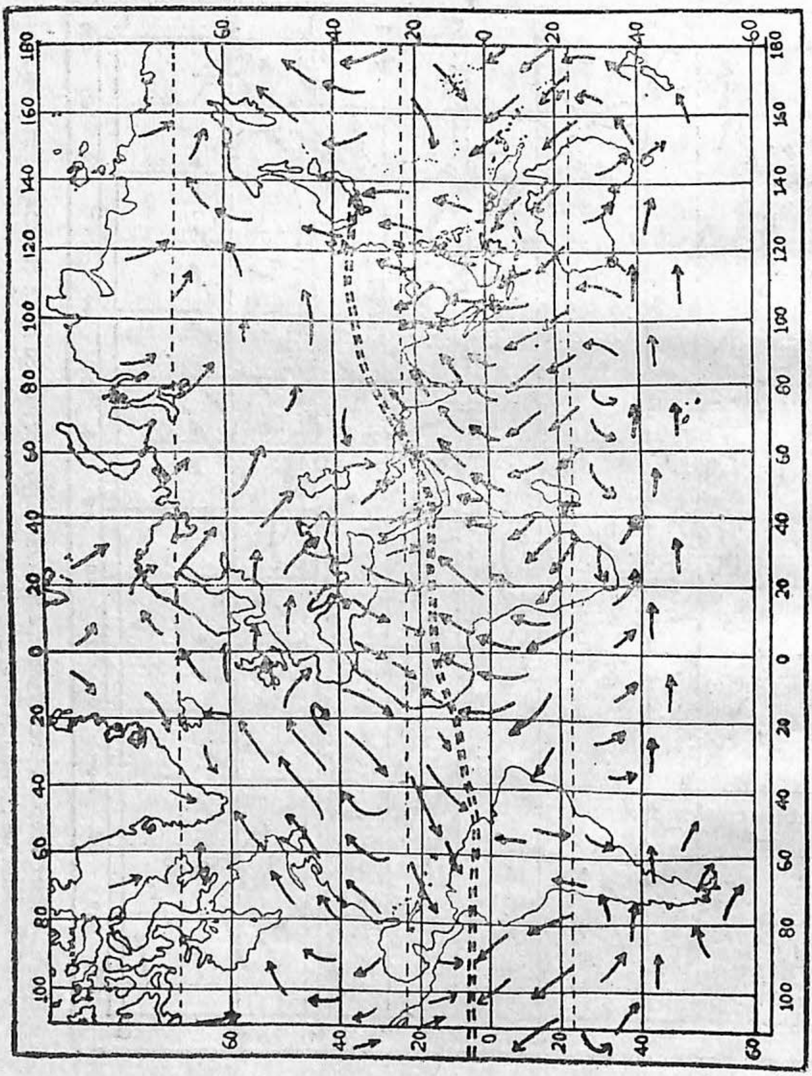

Fig. 1.2—Inter Tropical Convergence Zone (July) represented by thick broken lines. Arrows indicate prevailing wind directions.

maximum cloudiness. Recent data from weather satellites reveal that bands of maximum cloudiness do not always coincide with the ITCZ. On other hand, cloudiness seems to be associated with regions of low pressure on both sides of the equator. This has led some meteorologists to suggest "Near equatorial troughs" or "Near equatorial convergence lines" as alternative names for the ITCZ, but the position is by no means fully settled. It is not clear, for example, whether in reality we have two or more troughs near the equator, nor is it clear what happens to the ITCZ when the monsoon winds from the southern hemisphere penetrate into the northern hemisphere.

In figures 1.1 and 1.2 we show the approximate position of the Inter Tropical Convergence Zone and the main surface air streams for January and July. It is interesting to see how the air streams are deflected to the right as they cross the equator from the south. As would be explained later, this is a result of the earth's rotation from the west to the east. Figure 1.3 illustrates the monsoon regime of the earth according to the definition of C.S. Ramage.

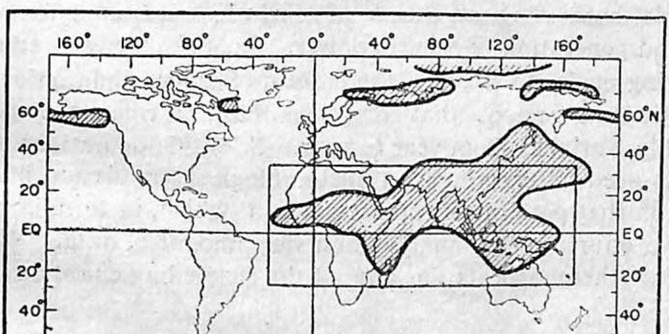

Fig. 1.3—The monsoon regime (from "Monsoon Meteorology" by C.S. Ramage, 1971). Hatched area is 'Monsoonal' and the rectangle broadly indicates the extent of the monsoon regime.

By far the most pronounced monsoon winds of the world are those that flow over the Indian sub-continent in the hundred-day period between the begining of June and middle of September each year. It brings relief to a dry and parched land in the form of rain, and affects Indian agriculture in a very substantial measure,

Indeed, the impact of the monsoon on Indian economy is very pronounced. It prompted a former Finance Member of the Government of India to say that his budget was 'largely a gamble on rain'.

The vagaries of the monsoon are now proverbial. The Indian farmer has had to put up with its temperamental nature on many occasions in the past. Excessive rain has led to floods in certain areas, while little or no rain in other parts has brought in its wake drought and famine resulting in acute distress to millions. Such fluctuations in rainfall have engaged the attention of our people and considerable effort has been directed from very early times to avert these calamities. There are many legends in our land of worship to the rain God for averting famines, and prayers are offered for pacifying the turbulent rivers of India. The Indian poets have sung about the rainy season in prose and verse, as though it formed an essential part of our national life and economy.

With the advent of modern technology, the emphasis has turned towards harnessing an excess of rain, where it occurs, for irrigation and generation of electric power. In a similar vein, attempts are being made to change the pattern of crops, and agricultural operations in regions that receive less rain, or where the rainfall is highly variable from year to year. Scientific operations of this nature need information on meteorological trends and periodicities in the pattern of monsoon rain which, in turn, implies a massive effort in the analysis of a vast amount of data. We will examine these aspects in one of the succeeding chapters of this book.

The earliest systematic study of rainfall data appeared in 1889. It was a valuable publication by H. F. Blanford entitled *The climates and weather of India, Shri Lanka and Burma.* Even though the observations available to him were meagre, a careful analysis by Blanford yielded much valuable information on the distribution of rainfall.

Blanford's work was extended by Sir Gilbert Walker, who gave up a fellowship at Trinity College in the University of Cambridge, to become the head of the Indian Meteorological Service from 1904 to 1921. Sir Gilbert relied on finding associations between a meteorological element, such as, rainfall, and some other

variable at an earlier time. He discovered, for example, an apparently significant relation between monsoon rains over the Indian peninsula and South American pressure in April and May. The discovery of similar associations between the monsoon and other meteorological variables in different parts of the world led him to devise a prediction formula for the season's total rainfall. We will discuss this in a later chapter on long-range weather prediction, but a word of caution is in order.

The search for an association between different meteorological events is an extremely difficult proposition. There are stringent statistical tests which need to be satisfied before one can be sure that a relationship does, in fact, exist. In the history of geophysical sciences, very few associations have been established which have a definite predictive value, or which are significant in a statistical sense.

Sir Gilbert Walker was an able mathematician who performed a prodigious volume of computations in his search for scientific causal relations. One often marvels at his patience and perseverence, especially when we recall that this was done at a time when mechanical aids, such as, computers, were difficult to come by. But, despite his scientific ingenuity and dextrous computations, there is considerable debate even today about the predictive value of Sir Gilbert's results. But, his vision and originality have been widely acknowledged. The southern oscillation was indeed a perceptive discovery.

The progress of Indian meteorology took a marked turn for the better with the coming of sounding balloons for probing the upper atmosphere. Investigations of the upper air began towards the end of the nineteenth century. The earliest probes were made with balloons, kites or a combination of a kite and a balloon known as the 'Kytoon'. Subsequently, developments were aimed at the construction of a suitable meteorograph for upper air research. By a meteorograph we mean a combination of sensing devices for monitoring the temperature, the pressure and the humidity of the air as the balloon ascends through the upper atmosphere.

Another event of significance to Indian meteorology was the introduction of the Dines Meteorograph in 1928. This meteorograph was originally developed by W.H. Dines in England. It

had a series of sensing devices which recorded their observations in the form of a permanent trace on a silvered plate. After an ascent was over and the meteorograph was recovered, the traces could be read with the help of a microscope. The success of these early ascents obviously depended on the meteorograph returning to earth intact, and its subsequent recovery.

Despite this handicap, it was surprising to find that a large number of meteorographs were in fact recovered. In India, many successful ascents with the Dines Meteorograph were made by G. Chatterji and his collaborators at Agra. The results of these ascents provided Indian meteorologists with the first glimpse of the upper atmosphere over India.

In more recent times, emphasis has been placed more and more on understanding the scientific principles which determine monsoon rain. The development of new observing techniques with the help of rockets, weather satellites and high altitude balloons is providing the meteorologist with a fund of new information. Along with probing devices, rapid strides have been made in our understanding of the physics of air flow. It is now possible, for example, to construct mathematical models which simulate the behaviour of large-scale air motion, such as the Indian monsoon. Developments in this field have been particularly impressive on account of improvements in computer technology. The solution of intricate mathematical problems is now feasible in a matter of seconds with the help of electronic computers. We will see later how modelling experiments are helping us to understand the monsoon.

Important though it is, its influence on agriculture forms only one aspect of the Indian monsoon. The analysis of rainfall has an increasingly important bearing on the work of hydraulic engineers, who operate the major river valley projects of modern India. Of considerable interest is a pioneer work by Professor P.C. Mahalanobis at the time of the Second World War. He made an extensive study of the rainfall, the run-off and other meteorological features of the river basins of Orissa. One particular result, which proved to be of great interest and practical utility, was a prediction formula for the level of the river Mahanadi at a place named Naraj. This formula was widely used to forecast the height of the river at Naraj for periods of 24 hours in advance. The entire

work is available in a classical paper by P.C. Mahalanobis on *Rain storms and river floods in Orissa* in 1940.

Last, but by no means the least are the effects of mountain barriers on the Indian monsoon. It is well-known that a large part of the heavy rain that falls on the west coast of India is caused by a topographic barrier in the shape of the Western Ghats. Of no less interest is the presence of another massive barrier—the Himalayan mountains—along our northern borders. The question may well be asked : What would happen to the monsoon if the Himalayan barrier was not there ? This question is not merely one of academic interest for in trying to answer it one ends up by asking a number of variations on the main theme. Would it be legitimate to assume, for example, that in the absence of the Himalayas the monsoon would proceed right up to Siberia and alter the climate there ? What would happen to the rain which we get over northern India because of the Himalayan barrier ? Could the climate be altered by removing mountain barriers, or by changing the rate at which the land gets heated in summer ? In trying to answer challenging questions of this nature we often discover many fascinating aspects of the Indian monsoon.

It cannot be claimed at this stage that all the answers are known or fully understood. But, in the chapters that follow we will try and explain the techniques which have been used by meteorologists to understand and, if possible, to explain why the monsoon behaves the way it does.

CHAPTER II

GLOBAL MONSOONS

WHEN WE CONSIDER A complex meteorological system such as the monsoons of the world, it is important to observe that we are dealing with several scales of motion. Opinions differ on what should be the dimensions of "macro" or "micro" scales of motion, but notwithstanding divergence of opinion it is convenient to separate the large or planetary scale monsoons from the smaller, but equally important, regional effects. It needs to be recognized that the two scales are strongly coupled, and the monsoon is really an end-product of interactions between motion on different scales.

We need to consider scales not only in space but also in time. For a three-dimensional atmosphere, it is necessary to specify, broadly, the horizontal and vertical dimensions. For this purpose, the classification of table 2.1 is convenient to adopt.

Table 2·1
Scales of motion

System	*Approximate dimensions*		
	Horizontal Scale (km)	*Vertical Scale (km)*	*Time (hours)*
1. Macroscale			
(a) Planetary waves	5000	10	200 to 400
(b) Synoptic perturbation	500 to 2000	10	100
2. Mesoscale	1 to 100	1 to 10	1 to 10
3. Microscale	Less than 1/10	Less than 1/100	1/10 to 1/100

In this chapter, the macroscale aspects will be considered. By this we mean the undermentioned facets :

—differential heating between land and oceans
--onset of the monsoon
—recession of the monsoon
—interannual variability, Walker and Hadley Cells
—teleconnections, the Southern Oscillation and El Nino.

One could perhaps add to this list by considering interhemispherical interactions, but as we will discuss this topic later it will be postponed to another chapter.

Differential heating

It was shown how differential heating between the sea and land could lead to a monsoonal circulation akin to a gigantic sea breeze. Let this be examined in more detail:

The amount of heat that is needed to warm a unit mass of a substance through 1°C is defined as its specific heat (C). It is often convenient to consider another quantity, namely, the heat capacity (H). This is the product of density (D) and specific heat (C). We have

$$H=D\times C$$

The unit of H is calories per cubic centimetre (cm^3) for every degree centigrade (°C), because it represents the heat needed to alter the temperature of a unit volume of the substance through 1°C.

Typical values of H for air, water and soil are provided in table 2.2.

Table 2·2
Heat capacity (H)
Unit : Cal/cm^3/°C

(i)	Air	0·003
(ii)	Water	1·0
(iii)	Soil	0·6

A striking feature is the large difference in the heat capacities of air and water. This indicates that, while it needs very little heat to warm a unit volume of air through 1°C, a much larger

quantity of heat is needed to warm an equivalent volume of water through the same temperature change.

This is important for understanding monsoons. Consider, for example, a piece of land surrounded by a sea as in fig. 2.1.

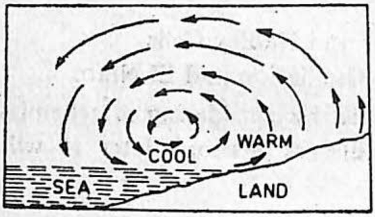

Fig. 2.1—Sea breeze—Day time cycle.

Both the land and the sea are warmed during the day by radiation from the sun. But, because solar radiation is only able to penetrate few centimetres of soil, the air above the land warms up much more rapidly because of the low heat capacity of air. On the other hand, the time needed to warm the sea is greater not only because water has much higher heat capacity, but also because solar radiation is capable of penetrating to a greater depth. As a consequence of this difference, the warm air rises over land and begins to spread out towards the sea. To compensate for the ascent of air over land a current of air is set up from the cooler sea to the land. Those who live near the sea will recognise this as the daytime sea breeze in the afternoon.

The situation is reversed at night because solar radiation is now cut off. The air above land cools very rapidly, again because air has small heat capacity, but the sea remains comparatively warm. This leads to a land breeze from a cooler land to the sea, that is, a wind opposite to the daytime sea breeze. The night cycle is shown in fig. 2.2.

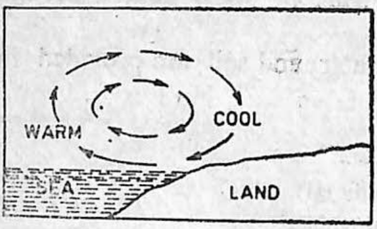

Fig. 2.2—Land breeze—Night time cycle.

Figures 2.1 and 2.2 illustrate how a reversal of winds is brought about by differential heating of land and sea. But this is motion on a small scale because the sea breeze seldom penetrates inland beyond the areas adjacent to the coast. Nevertheless, many features of similarity with the monsoon are readily discerned. In

summer, which corresponds to the daytime cycle of the sea breeze, monsoon winds blow from the sea towards the land. On the other hand, in the northern winter, which is similar to the night cycle (fig. 2.2), there are winds that blow from the land to the sea.

The monsoons represent winds on a much larger scale. This scale of motion extends over thousands of kilometres. Differential heating is still the principal driving force, but the path of air is now disturbed by diverse features, such as, the earth's rotation, mountain barriers and the retarding effect of friction as the winds blow over land. To this we should add interactions that take place between monsoonal currents and wind systems in higher latitudes, and between low level monsoons and systems in the higher reaches of the atmosphere. Another important difference lies in the condensation of water vapour within the monsoon air. This not only gives us rain, but is also accompanied by the release of large amounts of latent heat to the atmosphere.

The credit for recognising the importance of solar heating goes to Edmund Halley to whom a reference was made earlier. He wrote *An historical account of the Trade winds and Monsoons, observable in the Seas between and near the tropics, with an attempt to assign their physical cause.* This was published in the Philosophical Transactions of the Royal Society of London in 1686. In this paper Halley asked the question : "Why in the north part of the Indian Ocean the winds, which for one half year agree to those of the other two oceans, should change in the other half year, and blow from the opposite points, while the southern part of that ocean follows the general rule, and has perpetual winds about southeast"?

By the other two oceans, Halley was referring to the Atlantic and the Pacific. What perplexed him was that while the winds changed direction every six months over the north Indian Ocean, a similar reversal was not observed over the southern half of the Indian Ocean or over the Atlantic and the Pacific.

Halley sought to explain this by observing that "if a country lying near the sun prove to be flat, sandy, low land, such as the deserts of Libya are reported to be, the heat occasioned by the reflection of the sun beams, and its retention in the sand, is incredible to those that have not felt it; by which the air being

exceedingly rarefied, it is necessary that the cooler and more dense air should run thither to restore the equilibrium".

He then went on to write "it is to be considered that to the northwards of the Indian Ocean there is everywhere land within the usual limits of the latitude of 30°, viz., Arabia, Persia, India, etc. which for the same reason as the midland parts of Africa, are subject to excessive heats when the sun is to the north, passing nearly vertical; but yet are temperate enough when the sun is removed towards the other tropic".

The reversal of winds was ascribed by him to the movement of the sun and the consequent heating of the land. But, Halley was candid enough to acknowledge : "But I must confess that in this latter occurs a difficulty, not well to be accounted for, which is, why this change of the monsoons should be any more in this ocean, than in the same latitudes in the Ethiopic, where there is nothing more than a south-east wind all the year". Ancient geographers often alluded to the Southern Atlantic as the Ethiopic Ocean. This gigantic basin is separated by the continents of Africa and South America.

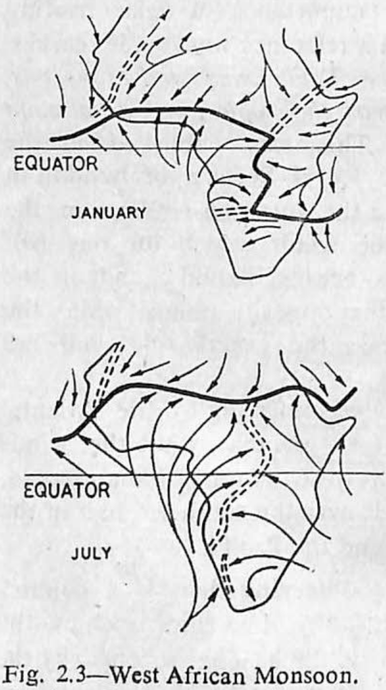

Fig. 2.3—West African Monsoon. Thick and dotted lines indicate air mass foundaries.

The question which Halley asked has not yet been satisfactorily settled, although we do know that there are other parts of the world where monsoonal winds occur. Apart from the summer and winter monsoons of Asia, parts of western and eastern Africa are influenced by a monsoonal type of circulation. Over west Africa, the south-east trades which flow from a region of high pressure around the island of Saint Helena (15°S, 5°W), cross the equator and, on being deflected by the earth's rotation strike the coast

of Guinea as a southwesterly wind (fig. 2.3). This monsoon is strongest in the summer months between June and September each year, but the rains commence in March-April after the sun moves northwards over the equator. As the sun moves southwards again the rains cease abruptly in October and November. The boundary between the southwesterly winds across the Gulf of Guinea and the northeasterly winds to the north is referred to as an Inter Tropical Discontinuity (ITD), while for the Indian monsoon we usually refer to it as an Inter Tropical Convergence Zone (ITCZ). There is a well marked southward shift in the location of the ITD over west Africa with the progressive movement of the sun to the south of the equator, but we do not observe a total reversal of wind direction from a southwesterly to a northeasterly direction, as is the case over the Indian subcontinent.

The movement of the Inter Tropical Convergence Zone (ITCZ) to locations north and south of the equator, in agreement with seasonal variations in the sun's angle of declination, provides the countries of eastern Africa with "long" and "short" rains. When the sun begins to move north after the winter solstice, the ITCZ is located over Kenya and Tanzania on most days of April and May each year. This wind discontinuity is responsible for much of the rainfall during these two months, which are referred to as the "long rains" of east Africa. Similarly, when the ITCZ begins its movement southwards, its location over eastern Africa provides a period of "short rains" in November. The interesting point to note is that while a difference in the response of land and ocean is responsible for the west African monsoon, seasonal variations of rain over east Africa are brought about by changes in the position of the ITCZ.

The question is sometimes asked : Why is the weather warmer in summer than in winter ? The answer to this is not because we are nearer to the sun in summer, as the earth is nearest to the sun around the first of January and farthest from the sun on the first of July. On an average, the earth is more than four million kilometers further from the sun in summer than in winter. The difference in solar energy received by the earth because of its varying distance from the sun is only seven per cent. This is insignificant.

The cause of summer and winter is to be found in the inclination of the earth's equator to the plane of its orbit around the sun. The orbit of the earth's revolution round the sun is in the form of an ellipse. The point at which the sun is nearest to the earth is its perihelion, while the aphelion defines the point at which the sun is farthest away. The plane of the earth's equator is inclined at an angle of $23\frac{1}{2}°$ to the plane of its orbit. This means that different parts of the earth receive varying amounts of sunlight as the earth revolves round the sun in an elliptic orbit (fig. 2.4). The northern

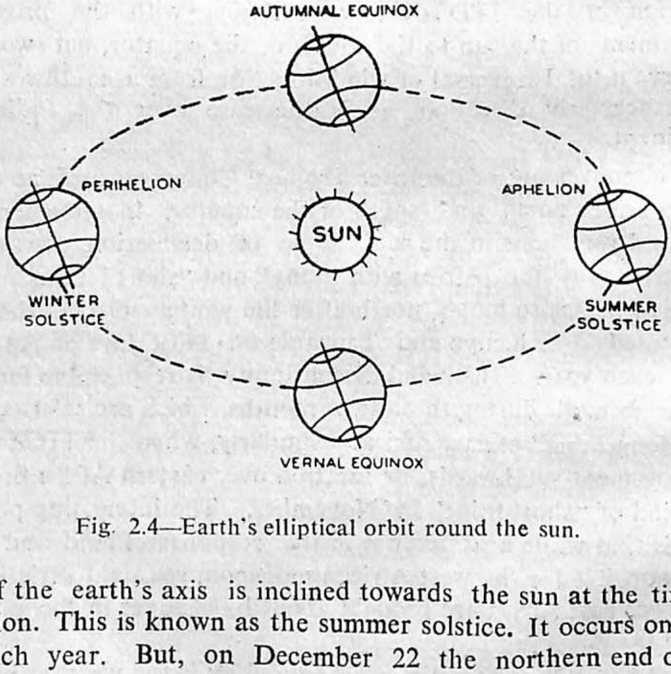

Fig. 2.4—Earth's elliptical orbit round the sun.

end of the earth's axis is inclined towards the sun at the time of aphelion. This is known as the summer solstice. It occurs on June 21 each year. But, on December 22 the northern end of the earth's axis is inclined away from the sun. This day is known as the winter solstice. In between the summer and winter solstices there are two days when the duration of sunshine is equal in both the northern and southern hemispheres. They are the autumnal equinox of September 22 and the vernal equinox of March 20.

If we assume that the earth is fixed in space, then to an observer stationed on the earth the sun and the stars would appear to be moving with "apparent motion". The plane in

which the sun appears to move is known as the plane of the ecliptic. This is the same as the plane of the earth's orbit. If we extend the earth's equatorial plane into space, we have what is known as a celestial equator. The equinoxes are located at the intersections of these two planes. The pressure pattern at the earth's surface during the northern summer is shown in fig. 2.5.

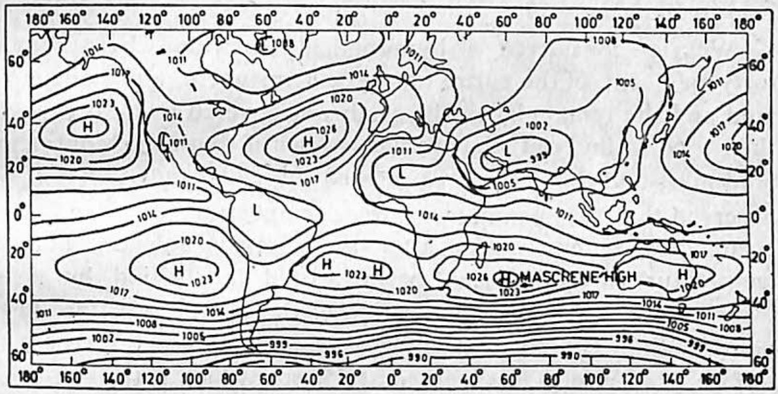

Fig. 2.5—Mean sea-level pressure in July. H—High Pressures ; L—Low Pressure. Figures indicate pressure in mb.

From this brief description of seasons, one may assume that, in the northern hemisphere, June should be the hottest month of the year and December should be the coolest, because they are the months which receive the largest and least amounts of radiation from the sun. But, usually there is a lag because time is needed for the heating and cooling to become effective. One notices, for example, the warmest time of the day is not at noon when the sun is overhead, but around two hours after the local noon. For the same reason the warmest months in the northern hemisphere are June, July and August, while the coolest months are December, January and February, that is, a little after the summer and winter solstice.

We can now see why the months of June, July and August are important for summer monsoons. They are the months when the tropics are warmest. But, it is important to realise that monsoonal winds depend on the thermal contrast between the

continents and oceans. The period when the continents are warmest need not of necessity represent maximum thermal contrasts because, as we can see, the oceans respond differently to solar radiation.

An interesting experiment was recently performed by P.J. Webster of Australia. His results have been published in the proceedings of a symposium on Monsoon Dynamics which was held in India in 1977.

Webster considered a hypothetical situation where roughly forty per cent of the entire northern hemisphere was made up of land and the remainder of the earth was covered by an ocean. He then tried to find out what would happen if both the continental land mass and the ocean was heated by solar radiation. It was observed that the maximum surface temperature over land would occur only a few weeks after the summer solstice, but the temperature in the adjacent ocean would lag behind by nearly two months. This is an indication of the slower response of the ocean. While the lag was clearly seen in temperatures near the surface, it was much less marked at 750 mb, which is the equivalent of an elevation of nearly of 2·5 km above the earth's surface.

For simplicity, Webster considered his continent to be land with uniform features, that is, there were no mountains or rivers in this experiment. This of course was unrealistic, but the important feature of the experiment lay in its success in simulating the broad features of monsoons. He found, for example, a low level southwesterly wind blowing from the sea to the land with an outflow of northeasterly winds aloft. This was similar to the summer monsoon over the Indian sub-continent.

Onset of the monsoon

A number of interesting changes occur in the circulation of the atmosphere when the summer monsoon sets in over India. The main features are :

(*i*) *The heat low*

As the sun moves northwards across the equator in the northern hemisphere, the continents surrounding the Arabian Sea begin to receive large amounts of heat; not only in the form of radiation from the sun but also as heat emitted from the earth's

surface. Professor Budyko from the USSR estimates that the flux of heat from the earth's surface into the atmosphere is the equivalent of 160 watts/m^2 for the month of June over the arid zones of northwest India, Pakistan, Saudi Arabia and the middle eastern countries. This is much larger than the corresponding figure of 15 watts/m^2 for the month of December, for example. Meteorologists measure the rate of input of heat energy by watts per square metre of the earth's surface (W/m^2). One watt for every square metre is roughly the equivalent of 62 calories for each square centimetre per month.

As a consequence of this large input of power, a trough of low pressure forms over this region. It extends from Somalia northwards across Arabia into Pakistan and northwest India. Towards the end of May, the heat low is well established and a southwesterly wind spreads northwards over the Arabian Sea, the Bay of Bengal and the Indian sub-continent (Fig 2.5). The onset of southwesterly winds over the west coast of India is often sudden; it is referred to as the "burst" of the monsoon, but there are other changes in the characteristics of the atmosphere before this happens. These will be discussed when regional aspects are considered.

(ii) Near-equatorial troughs and the Mascarenes High

Before the onset of the summer monsoon over India, a low pressure zone forms on either side of the equator, roughly along 5°N and 5°S. Meteorologists refer to this as an equatorial double trough. The double trough is frequently observed in satellite observations of clouds. Prior to the onset of the monsoon, the near-equatorial trough north of 5°N weakens, but the trough near 5°S remains fairly active. The view has been expressed that the equatorial trough along 5°N gradually moves northwards with the progress of the monsoon and, after merging with the heat low over Saudi Arabia, Pakistan and northwest India, it forms a quasi-stationary monsoon trough. On the other hand, some meteorologists feel the near-equatorial trough (5°N) and the monsoon trough are two different entities. This remains one of the unsolved problems of monsoon meteorology.

It has been suggested that the onset of the monsoon is related to a sudden acceleration of air from the southern hemisphere

towards India across the equator. The southern hemispherical circulation of the monsoon regime is dominated by an anticyclonic circulation, around a region of high pressure, off the coast of Madagascar. This is known as the Mascarenes high. In the southern hemisphere, the air blows round the Mascarenes high in an anti-clockwise direction. Opinion is still divided on what causes the air to accelerate round the Mascarenes high. Some feel this is brought about by the passage of migratory low pressure systems off the coast of South Africa, but this needs further investigation.

(*iii*) *Sub-tropical westerly and tropical easterly jet streams*

There are equally interesting changes that take place in the upper atmosphere with the advent of the summer monsoon. Towards the end of May, a narrow stream of air, which moves from the west to the east over northern India, suddenly weakens and moves to a new location far to the north of the Himalayas. This is known as a sub-tropical westerly jet stream. Its movement towards the north is one of the main features associated with the monsoon's onset over India. As the westerly jet moves north, yet another jet stream sets in over the southern half of the Indian peninsula. This flows in the reverse direction from the east to the west. It is called the tropical easterly jet, and it exhibits periodic movements to the north and south of its mean location during the hundred day monsoon season beginning with the first of June and ending around mid-September.

The altitude at which the winds attain their maximum strength in the tropical easterly jet is around 150 mb, but the maximum winds associated with the sub-tropical westerly jet occur at a lower altitude of 300 mb. A remarkable feature of the tropical easterly jet is that it can be traced in the upper troposphere right upto the west coast of Africa. The African monsoon is influenced by another jet stream, the African easterly jet, in the middle troposphere (700-600 mb) which has no counterpart in the Indian summer monsoon. An interesting feature of the African easterly jet is its association with the formation of waves in the middle troposphere. These waves have their origin around Khartoum

in east Africa and move westwards upto Dakar in Senegal on the west coast of Africa, before emerging into the Atlantic. The easterly waves are believed to be the precursors of hurricanes which eventually strike the Caribbean Islands.

(*iv*) *The Tibetan high*

A dominant feature, which is related to the monsoon's onset, is the emergence of a zone of high pressure over the plateau of Tibet. The reasons for the formation of the high are not very well understood at present. The average height of the plateau is around 4 km, which corresponds to a pressure of about 600 mb. This is nearly half the depth of the troposphere.

Chinese meteorologists have designed laboratory experiments to simulate the formation of this anticyclone. Their observations suggest widespread thunderstorms over the southeastern parts of Tibet in the pre-monsoon months of April and May. This releases considerable amounts of latent heat into the atmosphere through rainfall. Coupled with the flux of sensible heat from the plateau, the observations suggest about 145 watts/m^2 of power are injected into the atmosphere. This is only slightly less than the power associated with the heat low over northwest India and Pakistan, according to Professor Budyko's estimate. The plateau thus acts as an elevated "heat island". But, unlike the heat low over the Indian sub-continent, thermal convection is stronger over the Tibetan plateau because of its higher elevation. Consequently, while there is evidence of a heat low over Tibet at elevations around 500 mb (6 km), the ascending air rapidly spreads outwards both to the north and to the south of the plateau. The divergence of air leads to the formation of an anticyclone over Tibet around 300—200 mb (9 to 12 km). We will not reproduce the details of the laboratory model, but it is an interesting experiment which seems to explain why an anticyclone forms in the upper atmosphere over the Tibet plateau.

(*v*) *Changes in Kinetic energy*

A number of studies are now beginning to emerge which seek to quantify the energetics of the monsoon circulation. These studies have received considerable impetus from international data collection programmes. An International Indian Ocean Expedition

was organised between 1963 and 1966, which was followed by two experiments conducted, jointly, by India and the USSR in 1973 and 1977. More recently, a more intensive data collection effort was made under the aegis of another international experiment—the Monsoon Experiment or MONEX to give it its popular acronoym—in 1979. MONEX observations suggest a sudden and rapid increase in the kinetic energy of the winds over the Arabian Sea by an order of magnitude just prior to the onset of monsoon rains. This has been associated with the formation of a vortex off the coast of south India, but the mechanics of this vortex formation is a subject on which opinion is sharply divided. The onset vortex is observed in some years but not in others, while the increase in kinetic energy is a more general phenomenon.

The withdrawal of the monsoon

The change from a summer to a winter type of circulation with the retreat of the sun to the south of the equator is of shorter duration than the reversal from winter to summer. It is much more gradual than the abrupt changes associated with the onset of the summer monsoon. The monsoon begins to withdraw from northern India around mid-September. By the end of October it has usually withdrawn from the region north of 15 °N, and from Bangladesh and Burma. Finally, it withdraws from the extreme south of the Indian peninsula and Sri Lanka by December.

About the time of the monsoon's withdrawal, the subtropical westerly jet stream again reappears over the northwestern end of the Himalayas. Thereafter, it moves southwards to its usual location south of the Himalayas by the end of October. The easterly jet, which was a feature of the onset, disappears rapidly after the recession of the monsoon by early October. Many features that were associated with the onset phase of the monsoon now disappear, as the winter Asian monsoon begins to set in over southern India towards December.

Interannual variability, Walker and Hadley Cells

The principal monsoons of the world are the summer and winter monsoons of Asia and the monsoonal circulations over west and east Africa. Recent observations indicate that the lateral extent of these circulations is much larger than what was believed to be

the case a few years back. As our understanding improves, observations point to several features of similarity between the main monsoons of the world. These monsoons appear to be dominated by circulations that are either aligned in a north-south or an east-west direction. The rising branch of each circulation is located near a source of heat while the descending limb occurs over a heat sink. In honour of George Hadley and Sir Gilbert Walker these cells are now named Hadley and Walker Cells.

Hadley and Walker cells reveal a pattern in the atmosphere's response to differential heating. From the early years of the present century, physicists realised that the phenomenon of convection in a fluid often leads to the appearance of characteristic cells. H. Benard was able to demonstrate in 1901 that whenever a shallow layer of volatile fluid was cooled at its upper surface by evaporation and warmed from below, the entire fluid broke up into a number of separate cells. The cells had upward motion at the centre, outward diverging motion at the top and descending motion in the outer regions. Similar convection cells are not only observed in the laboratory, but also in the atmosphere and oceans. One has only to see cumulus clouds or vortices of different shape and size to realize the cellular structure of atmospheric convection.

It is still too early to assert whether Hadley and Walker Cells are in a sense the manifestation of cellular convection on a gigantic scale because the details remain to be investigated. But, if they are visualised as cells on a planetary scale then it is an interesting fact, because they form despite deviations brought about by the earth's rotation, topographic barriers and interactions between motion on different scales.

For the Asian summer monsoon, we have noted earlier that the plateau of Tibet acts as an elevated heat source. The ascending air above the source gradually spreads southwards to join a descending limb over the north Indian Ocean near the Mascarenes High. The southwesterly winds at the surface form the return current to complete the Hadley Cell. In addition, there is an east-west Walker Cell that appears to influence the summer monsoon. The ascending branch of the Walker Cell is located over Indonesia where, on account of convection and heavy precipitation, a heat source is generated. The descending limb of the

Walker Cell is located over the semi-arid regions of northwest India, Pakistan and the Middle East. It may appear at first sight surprising that semi-arid land, which one would normally expect to be a heat source should act as a heat sink. As explained in a later chapter on the Rajasthan desert, the high reflective power of the soil coupled with high temperature results in a larger amount of solar energy being lost to space compared to what is coming in from the top of the atmosphere. Consequently, deserts often act as sinks rather than sources of heat, and descending motion is needed to compensate the cooling by an excess of outgoing solar energy. The broad features of the Hadley and Walker Cells of the Asian summer monsoon are shown in fig. 2.6 (a).

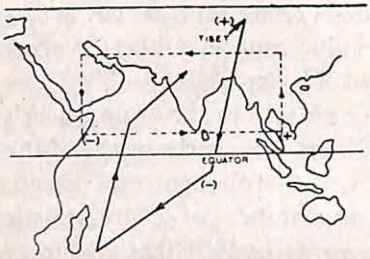

Fig. 2.6 (a)—Summer Monsoon Hardley Cell. (Broken line represents the Walker Cell). Ascending limbs are marked (+), while (—) denotes descending branch.

We will not proceed further at this stage with the structure of the Asian winter monsoon or the west African monsoons apart from stating that Hadley and Walker Cells also play a major role in other monsoon systems.

For the northern winter monsoon, the region over Indonesia and Malayasia acts as the main source of heat. This is again because of intense convection and large precipitation. The atmosphere receives a large amount of latent heat when water vapour condenses to form rain. The ascending branch of the Hadley Cell is thus located over Indonesia and as the ascending air spreads northwards it descends near an anticyclone over Siberia. The return current is in the form of cold surges from Siberia and adjoining China into Malayasia. These cold surges are often the precursors of spells of heavy rain over Malayasia during the winter monsoons.

It has been observed that several east-west circulations—Walker Cells—occur over region affected by monsoons. This is shown schematically in fig. 2.6 (b). It has been conjectured that the performance of the Asian summer monsoon is often deter-

mined by the relative importance of Hadley and Walker Cells. Good monsoons appear to be associated with more intense Hadley circulations and relatively weak Walker type of cells, while poor monsoons occur when the Walker Cell is strong and the Hadley

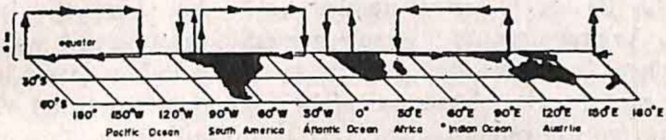

Fig. 2.6 (b)—Schematic diagram of Walker Circulations.

Cell is weak. The dynamics of this type of cellular convection—on a planetary scale—is often intimately linked with the interannual variability of monsoons.

Teleconnections, the Southern Oscillation and the El Nino

Towards the end of 1972 a series of catastrophic events in different parts of the world focussed attention on the possibility of global teleconnections in weather. The monsoon of 1972 was poor, and an unusual spell of lean rainfall over the Sahel region of northern Africa led to what is now known as the Sahelian drought.

Around the same time an abnormal current of warm waters off the coast of Peru in the eastern Pacific severely depleted the fishing industry of that country. Normally, the coastal waters off Peru and equatorial regions of South America are rich in sea food, especially in anchovies. But, once every five or ten years a warm current appears off the coast which reduces, dramatically, the yield of fish.

Such an event is called an El Nino. Literally, an El Nino means the 'Child of Christ', because it occurs around the time of Christmas. The El Nino of 1972-73 occurred at about the same time as a poor monsoon over India and drought over the Sahel. Were these events related or were they an accidental coincidence ?

In an attempt to answer questions of this nature there is now a resurgence of interest on teleconnections between abnormal weather in different parts of the world. Of considerable impor-

tance is a phenomenon called the "Southern Oscillation". This was discovered by Sir Gilbert Walker nearly sixty years ago. It postulates a see-saw pattern of weather between the Pacific Ocean and the Indian Ocean extending from Africa to Australia. Sir Gilbert discovered that when the pressures tended to be high over the Pacific Ocean it tended to be low over the Indian Ocean. As pressures are inversely related to rainfall, this suggests that when low pressures prevail over the Indian Ocean in the winter months, the chances are that the coming monsoon will be good in terms of rainfall. This fact is itill utilised in long range prediction of monsoon rainfall, which will be described in a subsequent chapter. The Southern Oscillation has a period varying from two to five years; consequently, pressure departures (deviations from the mean value) and their trends are better predictors than the absolute values of pressure.

In a sense the Southern Oscillation is another version of the Walker Cell. Many views have been advanced to explain this circulation. Dr. J. Bjerknes, a well-known meteorologist from Norway, explains the Walker Cell as a feed-back from the oceans. He suggests that the main drive for the Walker Cell comes from a difference in the temperature of the sea surface between Indonesia (warm) and the eastern Pacific Ocean (cold). Cold coastal waters of the eastern Pacific are generated by upwelling, that is, the replacement of surface waters by the colder water from greater depths. The Indonesian sector being warm in terms of sea surface temperature induces greater cloudiness which, in turn, cuts off the input of solar radiation, thereby reducing the driving for the Walker circulation. But, as the drive behind the Walker Cell weakens, the coastal waters off the eastern Pacific become warmer because of lesser upwelling. The sea surface temperatures rise, thereby reducing the temperature difference between the Indonesian sector and the east Pacific still further. In this manner, the Walker Cell passes from a phase of strong intensity to one of weak circulation until, as mentioned earlier, the pulsating nature of the circulation reverses the role of Indonesia and the eastern Pacific after two to five years. This is a broad and simple explanation in which many gaps remain to be filled in, but there appears to be an increasing body of evidence that points to the fact that convection over Indonesia and the eastern Pacific are

important for understanding the dynamics of the Asian summer and winter monsoons. The pressure difference between Tahiti in French Polynesia and Darwin in northern Australia is often taken to represent the intensity of the Southern Oscillation.

The above explanation, albeit in simple terms, suggests a close relationship between the El Nino, summer monsoon rain and the Southern Oscillation. An El Nino event in the preceding winter is suggestive of a strong Walker circulation and a weak monsoon. Although this did happen in 1972, it is necessary to sound a word of caution against premature inferences. A great deal of investigation, based on analysis of past data, remains to be done before firm conclusions are reached. For instance, if we examine the past history of Indian rainfall from 1875 to 1985, we will find about 43 years of deficient monsoon rain. But, only 19 of the 43 deficient years were El Nino years. This indicates that there are other factors, apart from an El Nino, which cause a deficiency in monsoon rainfall over India. There were only 6 years in the period 1875-1985, when an El Nino took place but rainfall was not deficient ; so it looks as though the El Nino does have some association with poor or indifferent monsoons, but the relationship is as clear-cut as some would have us believe.

CHAPTER III

REGIONAL ASPECTS OF THE MONSOON

IT IS CONVENIENT TO consider the regional aspects of the monsoon under two categories. In the first category, we discuss meteorological systems that generate short period variations in monsoon rainfall. By this we mean a transition from an active to a lean phase of the monsoon. A lean period of rain is usually referred to as a "break" in the monsoon. The usual duration of an active phase or a "break" is about a week, but on some occasions this could be longer. The longest breaks have been known to persist for two to three weeks, but such occasions are rare. In the second category, we will be interested in features that prevail throughout the monsoon, but whose dimensions are smaller than the planetary scales mentioned earlier.

Adopting this classification, we will discuss the features given in the following table.

Table 3.1

Regional aspects

I	II
Short period rainfall variations	*Regional scale features*
(i) Monsoon depressions in the Bay of Bengal	(i) The Somali current and allow level cross equatorial jet.
(ii) Mid-tropospheric low pressure systems	(ii) Air-sea interactions and temperature inversions.
(iii) Off-shore vortices	(iii) Movements of the monsoon trough.

We need to emphasise that the features under I and II are closely related. Moreover, one could view them as a part of world wide fluctuations occurring on a larger scale. Many of these interactions are not known very precisely, but they will be discussed qualitatively.

Short period rainfall variations

(*i*) *Monsoon Depressions*—A good part of the rainfall during the monsoon is generated by the westward passage of depressions or low pressure systems in the Bay of Bengal. On an average, one to three such systems are observed in the monsoon months, especially in July and August. The horizontal dismensions of a "low" or a depression are around 500 km.

The usual life time of these systems is around a week. Their mean tracks reveal west-north westward movement for the first three

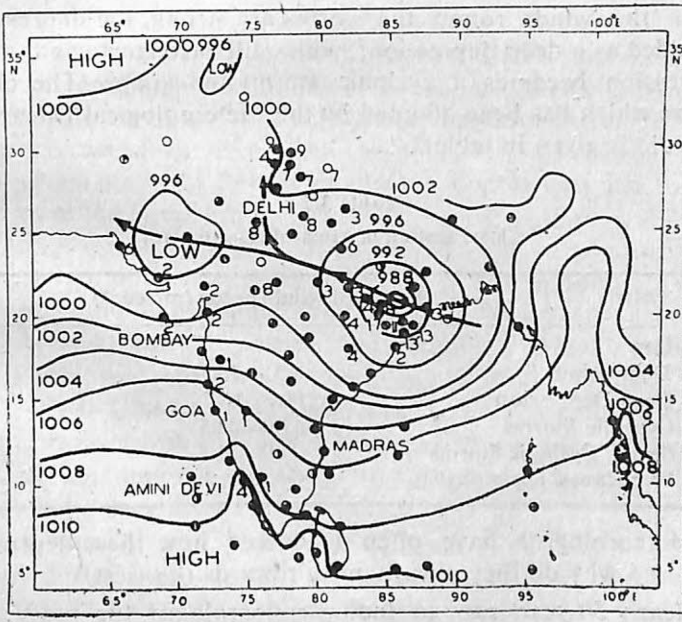

Fig. 3.1—Weather map of August 8, 1979 (0830 hrs. IST) showing surface pressure distribution and the 24-hour rainfall (cm) associated with a monsoon depression. Tracks followed by depressions are indicated by thick line.

to four days ; thereafter they tend to recurve north or northwest-wards. Figure 3.1 illustrates the surface pressure distribution and the 24 hour rainfall associated with a Bay depression, along with the mean track of depressions. There are deviations from the mean track for individual situations; some depressions do not recurve for example, but continue to move westwards and generate heavy rainfall over western India. A peculiar feature of these systems is the concentration of rainfall in the southwestern sector of the depression. Whenever a low or a depression forms, heavy rainfall is predicted all along the southern and southwestern sectors of its track. The structure of a Bay depression usually indicates a tilt in its vertical axis towards the southeast.

A depression or a low is an atmospheric vortex with a central region of low pressure. In the northern hemisphere, the winds blow round the centre in a counter-clockwise direction. The intensity of the vortex is measured by the strength of winds. Thus, when the winds round the vortex are strong, the depression is classified as a deep depression; with still stronger winds a deep depression becomes a cyclonic storm and so on. The classification which has been adopted by the Meteorological Department of India is given in table 3.2.

Table 3.2
Classification of atmospheric vortices

System	Range of wind speeds (m/sec^{-1})
Low	8.5
Depression	8.5 —13.5
Deep Depression	14.0—16.5
Cyclonic Storms	17.0—23.5
Severe Cyclonic Storms	24.0—31.5
Hurricanes/Typhoons	Greater than 32.0

Meteorologists have often wondered how these depressions form and why do they always move towards the west.

Nearly 50 years ago, an Indian meteorologist, the late V. Doraiswami Iyer, examined the tracks of cyclonic storms over the Pacific Ocean and the South China Sea for the years 1884-1930. He observed that 135 out of 370 cyclonic storms ultimately moved over India as residual low pressure systems. These systems traver-

sed the hilly terrain of Indo-China or South China and finally entered the northern part of the Bay of Bengal. This view did not find much support for many years because little data were available but it is now finding support through new observations. When we consider the fact that the late Mr. Iyer had very little by way of data, his early work was indeed commendable.

In a recent investigations two Indian meteorologists, K.R. Saha and J. Shukla collaborating with F. Sanders from the Massachussets Institute of Technology examined 52 lows and depressions in the Bay of Bengal during the ten-year period 1969-78. They observed that 50 of these disturbances were westward propagating systems from the northwestern parts of Thailand. Of this number, 32 were associated with a subsequent development in the Bay of Bengal. Sixteen of these disturbances were linked with the remnants of tropical cyclones, 14 with disturbances from South China Sea and 20 with disturbances that appeared to have had their origin over land.

Clearly, there is evidence to suggest that a fair number of monsoon depressions in the Bay of Bengal are activated by westward moving remnants of disturbances that have had their origin far to the east of the Bay. But there are also others who assert a direct relationship between monsoon depressions and tropical cyclones of the South China Sea.

They suggest that the formation of a Bay depression is the manifestation of a form of instability in the atmosphere, whenever the wind and temperature profiles favour the growth of instability. It can be shown on theoretical grounds that the atmosphere over the tropics could support several types of instability. By this we mean an atmospheric state which permits a small disturbance, in the form a propagating wave for example, to grow in amplitude and, finally, to form a closed circular vortex resembling a depression.

The atmosphere is defined to be in stable equilibrium if, after a small disturbance, it reverts back to its original state. In a stable atmosphere, air that is displaced upward or downward experiences forces which tend to restore it to its original level ; but in an unstable atmosphere the displaced air tends to continue to rise or fall, which ultimately leads to complete overturning. The stabilising or de-stabilising forces depend on whether the

disturbed air is warmer or colder than its environment which, in turn, depends on the vertical profile of temperature within the atmosphere.

Theoretical considerations indicate that if the decrease of temperature exceeds 1°C for every 100 metres, the atmosphere would be unstable to a disturbance that may be imposed on it. The rate of fall of temperature is known as the lapse rate to meteorologists. The critical lapse rate of 1°C per 100 metres is known as the dry adiabatic lapse rate. Physicists define a change as being adiabatic, if there is no exchange of heat when the change occurs. Thus, if a specified volume of dry air rises or falls without transfer of heat across its boundaries, its temperature changes would be dry adiabatic. A common example of an adiabatic change is the heating of a bicycle pump when it pumps air into its tyres. Here the pressure on a volume of air is suddenly increased, so that only a negligible amount of heat is conducted across the walls of the pump during compression.

Thunderstorms, tornadoes and other types of strong convection are associated with lapse rates that exceed the critical value of 1°C per 100 metres.

We have described here a simple form of vertical instability concerned with the up and down motion of a parcel of air. The stability of wind currents could be considered from another point of view. We may assume that a steady current of air exists in the atmosphere. If this is disturbed by a moving wave like disturbance, would the steady current allow the wave to grow? If it does permit growth, then we could state that the prevailing atmosphere was unstable; on the other hand, if it did not allow the disturbance to grow, it would be a stable atmosphere. This type of instability is referred to as hydrodynamic stability. Recent research suggests hydrodynamic instability could, on certain occasions, lead to depression formation.

In general, there is a close relationship between the temperature and wind profiles in the atmosphere, which we will discuss later. For the present, it is important to note that we must make a distinction between conditions that are necessary as well as sufficient for a depression to form.

What do we mean by necessary and sufficient conditions? This could be illustrated by the well known example of all the

three sides of an equilateral triangle being equal. The condition of having all its three sides equal is necessary for an equilateral triangle. Conversely, it may be demonstrated that if the three sides of a triangle are equal, it must be an equilateral one. The condition of three equal sides is thus both necessary and sufficient. Unfortunately, in many types of atmospheric circulations, it is possible to demonstrate that certain conditions are necessary, but are not sufficient by themselves. There is thus an inherent danger in picking upon stray examples and attempting wide generalities. That there is need for caution in this type of reasoning is illustrated by claims about the prediction of monsoon depressions in the Bay of Bengal. It has been claimed that if there was a decrease in the vector difference between the prevailing winds in the upper and lower atmosphere, then this would be a precursor for depression formation. It is easy to see why this condition should be necessary. Small differences in upper and lower atmospheric winds favour the development of convection and clouds, but this is not the case if large differences exist because, in the latter event, the tops of clouds would be sheared off. But is the vector difference between the upper and lower atmospheric winds sufficient ? It is clearly not, because it ignores the possibility of instability due to a favourable temperature profile. One cannot claim, therefore, that a decrease in wind shear would inevitably lead to depression formation. At best it represents a necessary condition but not a sufficient one.

In this manner, recent observations have suggested many indications on weather charts that favour the formation of a low or a depression in the Bay of Bengal. We have already mentioned a decrease in the vector difference between upper and lower atmospheric winds. A rapid fall in the surface pressure along the coastal stations in West Bengal, Orissa and north coastal Andhra Pradesh is another indication. A change in the direction of the prevailing winds at coastal stations is yet another favourable feature. If the winds tend to back with height thereby suggesting a change towards a counter-clockwise flow, it is usually taken to be an indication of depression formation. We would like to reiterate that these indications are in the nature of necessary conditions; they are not sufficient by themselves.

(ii) Mid-tropospheric disturbances—During the International Indian

Ocean Expedition in the late sixties, it was observed that very heavy rainfall over western India, especially over the northern parts of Maharashtra, Saurashtra-Kutch and Gujarat, was associated with cyclonic vortices that were confined to the middle atmosphere. They appear as circular vortices between 3 and 6 km, with their largest amplitude near 600 mb (4 km). The dimensions of these vortices are roughly of the order of 300 km in the horizontal direction and about 3 km in the vertical direction. A peculiar feature of these vortices is that they are only confined to the middle troposphere and are not visible at the surface. The structure of a mid-tropospheric cyclonic vortex reveals a core of warm air above the middle level and a slightly cold core below that level. Unlike the Bay of Bengal depressions, they show very little movement and appear to remain quasi-stationary for many days. The formation of these mid-tropospheric cyclones is often responsible for heavy rain on the northern sectors of the west coast of India. Rainfall amounts of 20 cm. in 24 hours are not uncommon. In fig. 3.2 we illustrate how a mid-tropospheric cyclone looks on the weather chart.

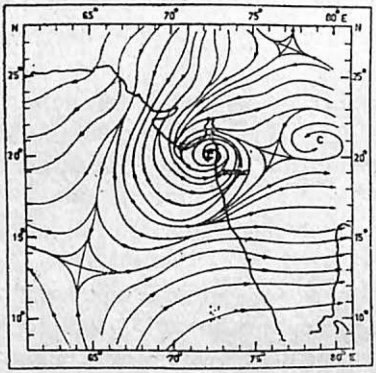

Fig. 3.2—Mid-tropospheric cyclone as seen on weather charts in the wind field at 4 km. above sea level.

Meteorologists have tried to explain the formation of a mid-tropospheric cyclone by examining conditions of instability. Studies reveal that if the winds in the lower level are stronger than those at the upper levels of the atmosphere, then it is possible to reproduce an unstable atmospheric mode which has some resemblance to a mid-tropospheric vortex. The precise formation of this type of vortex has not yet been satisfactorily explained.

(iii) Off-shore vortices—During the monsoon, we often observe spells of very heavy rain along the west coast of India. Apart from mid-tropospheric cyclones, these spells of heavy rain are

associated with off-shore vortices.. It is to be noted that the west coast has an orographic barrier in the form of the Western Ghats. The average altitude of the Western Ghats is between 1.0 and 1.5 km. These mountains run in a north-south direction, and are approximately 1000 km in length and 200 km in breadth. When the monsoon winds strike the mountains, on many occasions they do not have enough energy to climb over the Western Ghats. On such occasions, they tend to be deflected round the mountain and the return current forms an off-shore vortex. These vortices have very small linear dimensions. Their diameter is only of the order of 100 km and their presence is often detected by a weak easterly wind at coastal stations. Notwithstanding their small dimension, they are capable of generating spells of very heavy rain lasting for about 2-3 days during the monsoon season.

Regional scale features

(i) Low level cross-equatorial jet and the Somali current—During the late sixties and early seventies, J. Findlater, a British meteorologist who was then stationed in Kenya, observed very strong winds in the form of a narrow current of air off the coast of East Africa. This low level jet stream was found to be most pronounced between 1.0 and 1.5 km. The general structure of the low level jet stream is shown in figure 3.3. It was observed to flow from Mauritius and the northern part of the island of Madagascar before reaching the coast of Kenya at about 3°S. Subsequently, it ran over the plains of Kenya, Ethiopia and Somalia before reaching the coast again a round 9°N. The jet stream appears to be fed by a stream of air which moves northwards from the Mozambique Channel.

The major part of the low level jet penetrates into East Africa during May and, subsequently, traverses the northern parts of the Arabian Sea before reaching India in June. Observations suggest that the strongest cross-equatorial flow from the southern to the northern hemisphere during the Asian summer monsoon is in the region of the low level jet. This has intrigued meteorologists, because it is not clear why the major flow of air from the southern to the northern hemisphere should take place along a narrow preferred zone off the East African coast. The importance of the low level jet also arises from the fact that its path around 9°N

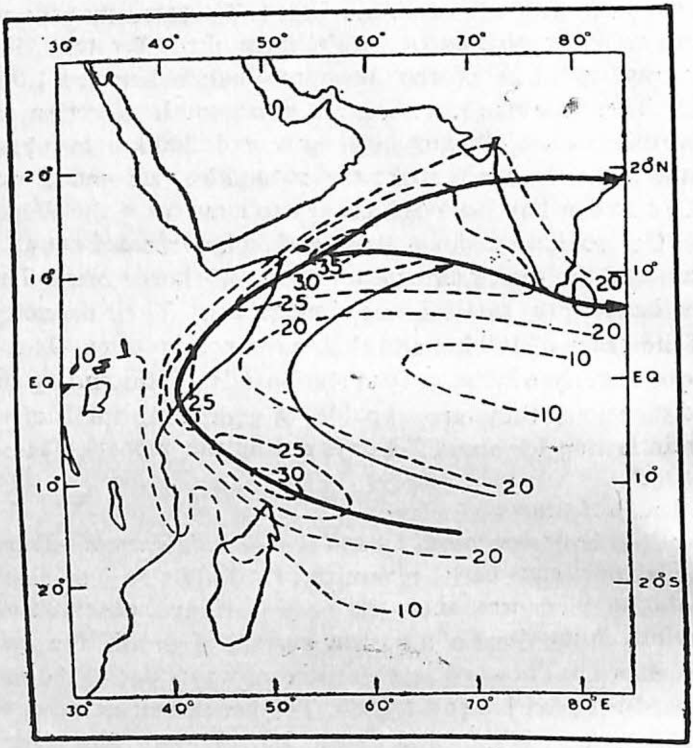

Fig. 3.3—General structure of low-level Jet Stream at 3000 ft (1 km.) in the month of July (after Findlater). Dotted lines show wind speed in knots.

coincides with a zone of coastal upwelling. As the strong winds drive away the surface coastal waters towards the east, extremely cold water from the depths of the sea rise upwards to preserve continuity of mass. This upwelling is brought about by strong low level winds.

Coastal upwelling is responsible for creating a narrow strip of cold sea surface temperatures off Somalia. The sea surface temperature in this region could be as low as 15°C, while off Bombay on the west cost of India the sea surface temperature is between 25–30°C. The strong temperature gradient that is set up has an impact on monsoonal winds.

When the sea is warmer than the overlying atmosphere, it

tends to supply heat to the atmosphere, but when it is colder it extracts heat from the atmosphere. Consequently, the sea surface temperature regimes that are created by the equatorial jet tend to generate two regions within the Arabian Sea, one which supplies heat to the atmosphere and another which draws upon the heat energy of the atmosphere.

After the low level jet moves towards the Indian coastline around 9°N it separates into two branches. One appears to move to the northern parts of the Indian Peninsula while the other recurves towards the southern half of the Indian coastline and Sri Lanka. It is still not very clear why the jet separates into two branches. Findlater analysed the wind profile for the months of July and August and found a relationship between the cross equatorial air flow, between 1.0 and 1.5 km, over Kenya and rainfall over western India. He was of the view that an increase in the cross-equatorial flow was followed by an increase in rainfall over the west coast.

The winds tend to flow across the lines of equal pressure (isobars) at the entrance region of the jet, just off the northern tip of Madagascar. This suggests rapid acceleration of the air as it enters the jet. The is a corresponding deceleration at the exit of the jet around 9°N.

Oceanographers have been interested in yet another phenomenon which appears to have some relationship with the low level jet stream off the coast of eastern Africa. This is an ocean current named the Somali Current, which flows northward from the equator to about 9°N, where it separates from the coast. It is a fairly strong current with a velocity maximum of 2 m sec^{-1}, but speeds as large as 3 m sec^{-1} have been also observed.

Over many oceans of the world, one observes an intensification near their western boundaries. Examples of such intensification are the Gulf Stream on the western boundary of the Atlantic Ocean and the Kuroshio Current on the western boundary of the Pacific. These currents are usually too strong to be treated as merely an oceanic response to local winds. Professor Henry Stommel, a well known Oceanographer, pointed out that western intensification could be partly accounted for by the latitudinal variation of the earth's rotation.

The Somali Current may be considered to be a western boundary current of the Indian Ocean. But, its peculiar feature is a reversal in direction with the onset of the summer monsoon. In winter, this current is from the north to the south running southwards from the coast of Arabia to the east African coastline; but with the advent of the summer monsoon it reverses its direction and flows from the south to the north. This suggests a relationship with the reversal of monsoon winds, but usually the oceans respond very slowly to changes in atmospheric circulation, and oceanographers have wondered why the Somali Current reverses its direction and reaches its maximum speed nearly a month earlier than the onset of southwesterly monsoon winds. In a celebrated paper on this subject in 1969, Professor Lighthill, a British mathematician, was able to demonstrate that a faster response could indeed be achieved if we took into account the vertical structure of the Somali Current.

(*ii*) *Air-sea interactions and temperature inversions*—We mentioned earlier that extremely cold sea surface temperatures prevail over the western sector of the Arabian Sea. This generates an atmospheric state where a layer of warmer air lies over the colder air beneath. Meteorologists refer to this as a temperature or thermal inversion, because the lapse rate which is usually negative becomes positive when warm air lies over colder air.

In view of the cold sea surface temperatures off the coast of East Africa, there is a pronounced temperature inversion in the atmosphere in this region. The base of the inversion is around 1.0 to 1.5 km above the sea surface. The height of the inversion base increases as we proceed eastwards and becomes gradually less marked. The inversion usually disappears as we move eastwards of 55°E. It inhibits formation of convective activity to the west of 65°E, and usually clear skies are observed on the western sector of the Arabian Sea. But, eastwards of 65°E the inhibiting effect of the inversion wears off and cloud development takes place rapidly. The convective clouds are often accompanied by heavy rain. The inversion layer is thus a measure of the modification of monsoon air as it traverses the Arabian Sea.

(*iii*) *Movement of the Monsoon Trough*—Mention has been made of a zone of low pressure which builds up over northwest India as

a result of excessive solar insolation. With the advance of the monsoon, this heat low gradually extends eastwards until it forms an elongated low pressure zone running parallel to the Himalayan mountains in a west to east direction. Meteorologists in India refer to this as the monsoon trough. Its axis is roughly parallel to the Himalayan foothills. The normal position of the monsoon trough is shown in fig. 3.4.

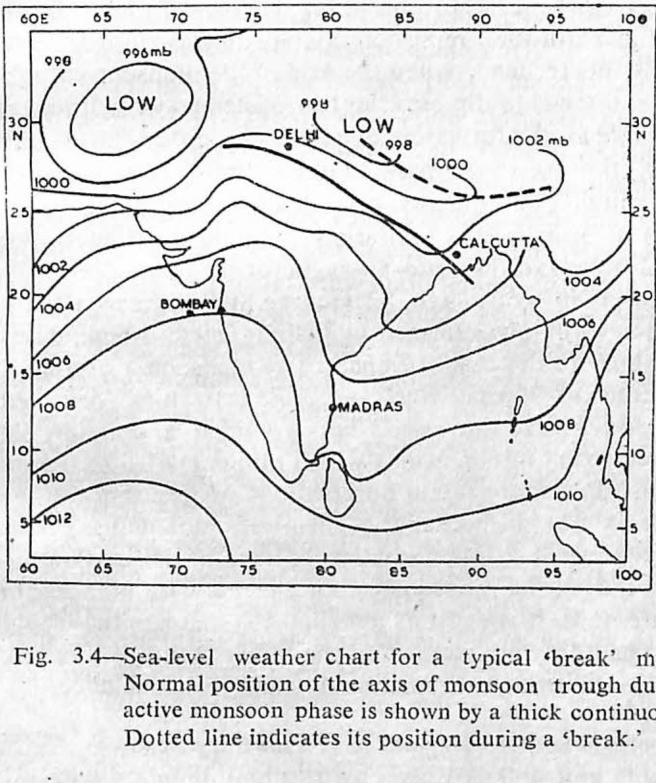

Fig. 3.4—Sea-level weather chart for a typical 'break' monsoon. Normal position of the axis of monsoon trough during an active monsoon phase is shown by a thick continuous line. Dotted line indicates its position during a 'break.'

The monsoon trough is not a quasi-stationary system. It shows periodical movements to the north and south of its normal position. When it moves north and lies close to the Himalayan foothills, there is a remarkable change in the rainfall pattern over India. The rains cease abruptly over the plains of northern India, but increase equally rapidly in intensity over the foothills of northeast India. This is known as a "break" in monsoon rains.

It leads to a paradoxical situation when people in the plains complain about lack of rainfall, while those living in the northeastern parts of our country are distressed by floods, because most of the major river systems of India have their origin in the Himalayan region. The weather chart for a typical "break" monsoon is depicted in fig 3.4.

Another feature of a "break" is the westward passage of low pressure systems across the Indian peninsula. Parts of the peninsula which lie in the rain shadow of the Western Ghats derive much of their monsoon rainfall during "break" situations. On the other hand, when the axis of the monsoon trough moves south and tends to dip into the Bay of Bengal, conditions become favourable for the formation of a low or a depression. As mentioned earlier, the westward passage of a depression is accompanied by heavy rain. Consequently, a southward position of the monsoon trough is usually an indication of well distributed rain over central India and the Indo-Gangetic plains.

Nearly 60 years ago, Sir George Simpson suggested that the monsoon trough was formed by the alignment of mountains to the north and to the east of India. The monsoon air on hitting the mountains of Burma were, according to him, deflected to the north and then to the west. This provided a counter clockwise rotation to monsoon winds which ultimately led to the monsoon trough. But, more resent observations suggest that the monsoon trough is not a mechanical effect of mountains. It is more closely related to the radiation balance of the earth-atmosphere system during the monsoon. Experiments with numerical models indicate that it was not possible to simulate the trough if the model had only mountains; on the other hand, it was possible to simulate the trough if radiational balance was included.

Interactions between Regional and Planetary Scales

We will end this chapter by touching upon an aspect which is now engaging the attention of research scientists in many parts of the world. This concerns interactions between the regional and planetary scales of the monsoon. Several view points have been put forward in recent years about different types of interactions, but, as of today, we find it difficult to discern firm results. Consequently, our intention here is to only indicate the possibility of interactions on which some evidence is beginning to emerge.

But, to reiterate, the whole subject needs more examination and the technical details are outside the scope of this book.

The first type of interaction is with systems that propagate eastwards in mid-latitudes to the north of India, especially with wave-like disturbances that move from the west to the east between 40° to 60°N. Some of these waves are of large amplitude. Many feel that an interaction of this type could lead to weakening of the upper tropospheric anticyclonic circulation over India at an elevation of around 9.0 km during the monsoon. Evidence has been subsequently put forward to suggest that when this happens, conditions favour a "break" in monsoon rain, but we do not yet understand whether this type of interaction does, in fact, take place, because data over the Himalayas are difficult to come by. This opinion has been expressed by others to suggest that the monsoon is a closed circulation with very little penetration from latitudes to the north of India because of the presence of Himalayas. Theory suggests that interactions between mid-latitudes and the tropics are only possible at certain preferred times of the year, mainly around the time of the equinoxes. Consequently, the evidence of mid-latitude systems having an impact on monsoons is not yet conclusive.

Mention has been made of the quasi-biennial oscillation in the equatorial upper atmosphere. The quasi-biennial oscillation is a change over from an easterly to a westerly wind regime in the equatorial stratosphere once every 24-30 months. This oscillation is believed to be an outcome of waves propagating vertically from the troposphere to the stratosphere. Evidence is being collected to ascertain whether the quasi-biennial oscillation has an impact on rainfall and sea surface temperature, for example. If this relationship could be established, it would have interesting consequences for long range prediction of monsoon rainfall. One could then see if there are rainfall variations between the easterly and the westerly phases of the stratospheric quasi-biennial oscillation.

As we have noted earlier, the monsoonal flow in the lower atmosphere below 6.0 km undergoes pronounced changes during active and "break" phases of the monsoon. How far are these changes influenced by events that take place far to the east or to the west of the monsoon regime is still a matter of conjecture.

There is considerable evidence now to suggest that the atmosphere could support a large number of waves on different scales. Some evidence has been produced to suggest two main periods of oscillations in monsoon rainfall : one is of five days duration and another with a period of 15 days. They suggest a rhythmic fluctuation in the intensity of rainfall with a period of 5 or 15 days. How far are these oscillations linked with the monsoon regime is yet another problem concerning interactions. As more and more data begin to come in, and as meteorologists begin to acquire better facilities for observing sequential changes in weather, we expect to improve our understanding of these features in the years to come. At present the evidence of such periodicities is not conclusive

CHAPTER IV

MONSOON CLIMATOLOGY

Dates of onset and retreat

The Asian summer monsoon spreads its influence over many countries of east and southeast Asia. In China, the summer rains are known as Mai-Yu, in the southern parts of Korea it is called Mae-ue, while in Japan the rains of early summer are popularly known as the Baiu rains. The summer rains of Japan are closely related to the arrival of the southwest monsoon.

Table 4.1 provides approximate dates for the beginning of the rainy season. It covers a region extending from 75 to 140° E and 20 to 45°N. The rainy season begins earliest in south China towards the beginning of May. Subsequently, it gradually extends its influence over the southeastern parts of Japan. Over India, the monsoon rains begin towards the very end of May or the first week of June over the extreme southern parts of the peninsula. The areas under thick lines in table 4.1 indicate regions where the rainy season begins in May or the first week of June. It is interesting to note that these dates which indicate the beginning of rains do not of necessity coincide with the onset of monsoon winds.

Opinions differ on what really constitutes an onset of the monsoon. Often it has been thought of as a spectacular event accompanied by thundersqualls and thundershowers. More often than not, this is not the case. The monsoon is basically an air stream laden with moisture. Its arrival is a gradual process beginning with a short period of transition from extreme heat to a very humid atmosphere with light rain. There are large areas in India, such as Assam and West Bengal, which experience violent thunderstorms in the pre-monsoon months of April and May. These are

Table 4.1

	75°E	80	85	90	95	100	105	110	115	120	125	130	135	140°E
45°N	—	—	—	—	—	—	—	—	—	Aug. 9-13	July 15-19	July 15-19	—	—
40	—	—	—	—	—	—	—	July 25-24	July 20-24	July 20-24	July 15-19	—	—	June 25-29
35	—	—	—	—	—	—	July 20-24	July 20-24	July 15-19	July 10-14	July 10-14	June 20-24	June 15-19	June 15-19
30	June 30 -July 4	June 15-19	June 10-14	May 31 -June 4	—	June 5-9	June 25-29	June 20-24	June 15-19	June 15-19	June 15-19	June 5-9	May 26-30	May 21-25
25	June 15-19	June 10-14	June 10-14	June 5-9	May 31 -June 4	May 31 -June 4	May 31 -June 4	June 10-14	June 5-9	May 30 -June 4	—	May 15-20	—	—
22.5	June 10-14	June 10-14	June 10-14	June 5-9	May 31 -June 4	May 21-25	May 21-25	May 16-10	May 6-10	May 6-10	—	—	—	—
20°N	June 10-14	June 10-14	June 5-9	May 26-30	May 26-30	May 21-25	May 1-5	May 1-5						

Table 4.1 : Beginning of the rainy season in monsoon regions of Asia from "Water Balance of Monsoon Asia" by M M. Yoshino. Thick (—) lines indicate regions where rains commence earlier, in May or first week of June.

the well-known "Nor' Westers" or "Kal Baisakhis" of northeast India. These thunderstorms are examples of intensive atmospheric vortices of small dimensions. They are associated with strong convective motion. On many occasions they appear as a group of cells, or large shower clouds, which are capable of releasing considerable amounts of precipitation. In some years, when pre-monsoon thunderstorm activity is particularly severe and frequent, it is difficult to distinguish between a pre-monsoon thunder-shower and genuine monsoon rain. But, on most occasions there is a fairly well defined period of transition during which atmospheric characteristics reveal a gradual change.

On a statistical basis, there is a method of delineating the normal dates for onset of the monsoon, which is convenient and satifactory. If we divide a month into 5-days periods or pentads as they are generally called, it is possible to compute, on a statistical basis, the normal rainfall for each pentad from the climatological records. It is possible to do this for all parts of the country over which a fairly dense network of raingauges has been in existence for over 75 years. Around the time of arrival of the monsoon, the normal rainfall over a particular pentad exhibits a sudden and well marked rise over its two or three preceding pentads. The increase is not merely transient, but is generally maintained. In these circumstances, we have no difficulty in stating that the normal date of arrival of the monsoon for a particular pentad which exhibits a sudden increase in rainfall. On this basis, the normal dates of onset of the monsoon have been fixed for different parts of India which are shown in figure 4.1.

The following working rules are often used to determine the date of onset over the southern tip of India :

(*i*) Beginning with May 10, if at least five out of ten meteorological stations in Kerala report 24 hour rainfall amounts of 1 mm or more for two consecutive days, the monsoon's onset is declared on the second day.

(*ii*) If three or more out of seven stations in Kerala report no rainfall for the next three days, a temporary recession of the monsoon is indicated. A temporary recession is not unusual when the monsoon is still south of 13°N.

(*iii*) After the monsoon has set in north of 13°N, it is taken as established over Kerala.

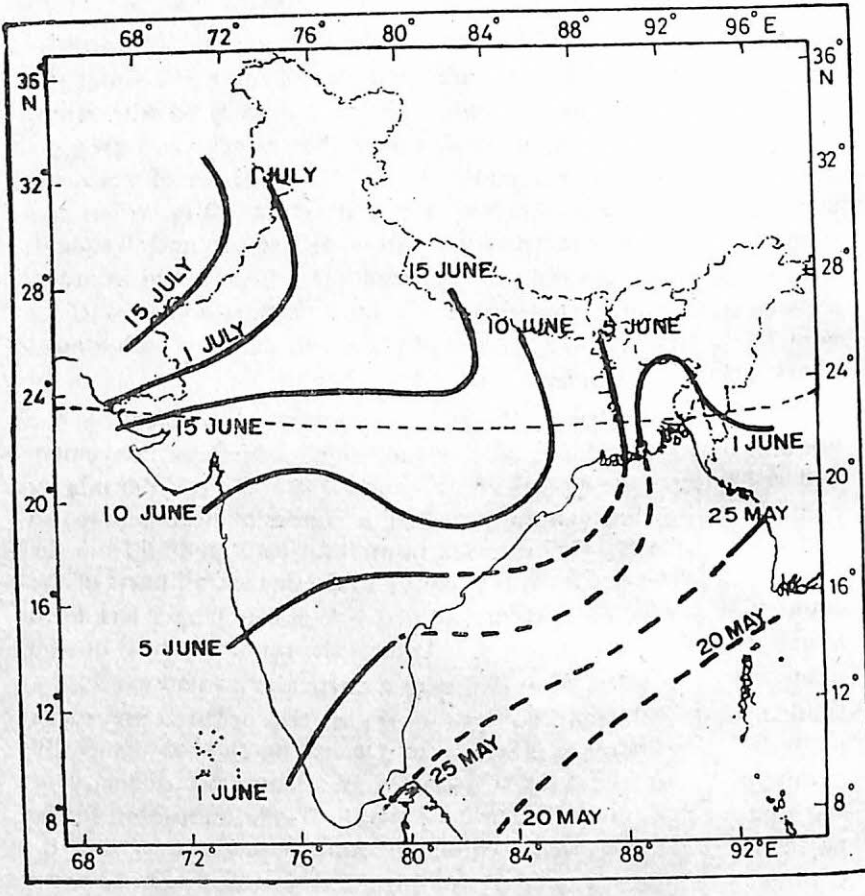

Fig. 4.1—Normal dates of onset of monsoon.
The territorial waters of India extend into the sea to a distance of twelve nautical miles measured from the appropriate base line. Responsibility for the correctness of internal details shown on the map rests with the publisher.

Similar working rules are used by meteorologists in the other states of India.

It will be observed from fig. 4.1 that the normal date of arrival of the monsoon in Sri Lanka and over the islands in the Bay of Bengal is towards the last week of May. Thereafter, it reaches the extreme south of the Indian peninsula a week later (June 1). The subsequent progress of the monsoon may be conveniently traced in the form of two branches, namely, the Arabian

Sea branch and the Bay of Bengal branch. The Arabian Sea branch gradually advances northwards to Bombay by June 10. The advance from Trivandrum to Bombay is achieved in about ten days and is fairly rapid.

In the meantime, the progress of the Bay of Bengal branch is no less spectacular. It moves northwards into the central Bay of Bengal and rapidly spreads over most of Assam by the first week of June. On reaching the southern periphery of the Himalayan barrier, the Bay branch of the monsoon is deflected westwards. As a consequence, its further progress is towards the Gangetic plains of India rather than towards Burma. The arrival of the monsoon over Calcutta is slightly earlier than at Bombay. The normal date of arrival at Calcutta is June 7, whereas the Arabian Sea branch of the monsoon normally strikes Bombay three days later, that is, on June 10. By mid-June the Arabian Sea branch spreads over Saurashtra-Kutch and the central parts of the country. Thereafter, the deflected current from the Bay of Bengal and the Arabian Sea branch of the monsoon tend to merge into a single current. The remaining parts of west Uttar Pradesh, Haryana, Punjab, eastern half of Rajasthan experience their first monsoon showers by the first of July. The arrival of the monsoon at a place like Delhi (28°N, 77°E) often raises an interesting question. Sometimes, the first monsoon showers at Delhi arrive from the east as an extension of the Bay of Bengal branch but on a number of other occasions, the monsoon is ushered in from the south, that is, from the Arabian Sea. The meteorologist is often confronted with the problem of trying to decide whether the monsoon will strike Delhi from the east or from the south.

By mid-July, the monsoon extends into Kashmir and the remaining parts of the country but only as a feeble current because, by this time, it has shed most of its moisture.

The normal duration of the monsoon is roughly 100 days beginning from June 1. It begins to withdraw from Punjab and Rajasthan by the middle of September. The withdrawal of the monsoon is a more gradual process than its onset. In general terms, the monsoon withdraws from the northwest India by the end of October and from the remaining parts of the country by early December. Fig. 4.2 indicates normal dates for the withdra-

wal of the monsoon. It is difficult to decide when the southwest monsoon ends and when the northeast or winter monsoon begins over the extreme south of the Indian peninsula. The winter (northeast) monsoon sets in over the southern half of the Indian

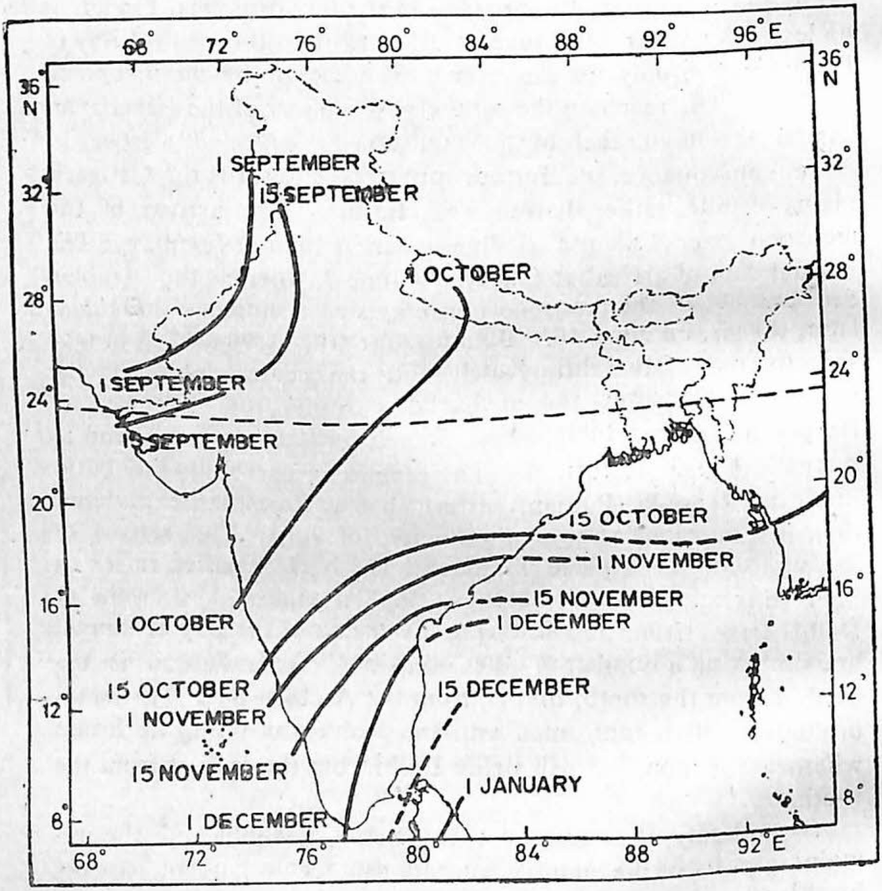

Fig. 4.2—Normal dates of withdrawal of monsoon.
The territorial waters of India extend into the sea to a distance of twelve nautical miles measured from the appropriate base line. Responsibility for the correctness of internal details shown on the map rests with the publisher.

peninsula in October. Although it may seem theoretically possible for both monsoon systems to exist over peninsular India, in reality

such situations are comparatively rare. The southwest monsoon is for all practical purposes in the last stages of its life towards the end of October. Rainfall in the southern states of Tamil Nadu, Karnataka and Kerala during November and December is usually attributed to the winter or northeast monsoon.

Monsoon rainfall

Over 70% of the annual rainfall over India is recorded during the southwest monsoon. The regions which receive the largest rainfall are along the west coast of India and the states of Assam and West Bengal in northeast India. In these regions orographic features play an important role, because moisture laden monsoon winds strike against physical barriers by way of mountains. The distribution of monsoon rainfall over India is shown in fig. 4.3.

Along the west coast of India the orientation of the Western Ghats is from the north to south. Normally, the monsoon winds which strike the Western Ghats from a southwesterly direction, shed most of their moisture on the windward side of the Western Ghats. Thus, while Bombay (18° 54′ N, 72° 49′ E) records about 187.5 cm of rain during the monsoon, Pune (18° 32′ N, 73° 70′ E) to the lee of the—Ghats receives only 50 cm of rain during the same period. Yet the distance between Bombay and Pune is only 160 km.

Perhaps the most well-known feature of orographic rain in northeast India is the extremely large amount of rainfall at Cherrapunji (25° 15′N, 91° 44′E). This small town in Meghalaya records an annual rainfall of about 1087 cm (428 inches) of rain. Of this amount, about 254 cm (100 inches) falls in the months of June and July. Approximately 65% of the yearly rain is recorded in the monsoon months of June, July and August, and only a very small part occurs in December and Jaunary. On June 14, 1876, Dr. Blanford, who was then the Chief of the Indian Meteorological Service, reported 103·6 cm (40·8 inches) of rain in Cherrapunji on a single day. The rainfall at Cherrapunji is believed to be the highest ever recorded in the world, although some claim that the record should now go to another nearby village Mawsynaram (25° . 18′ N, 91° . 35′ E). The Mawsynaram records are not so reliable as those of Cherrapunji, but the available data suggest that the mean annual rainfall at Mawsynaram

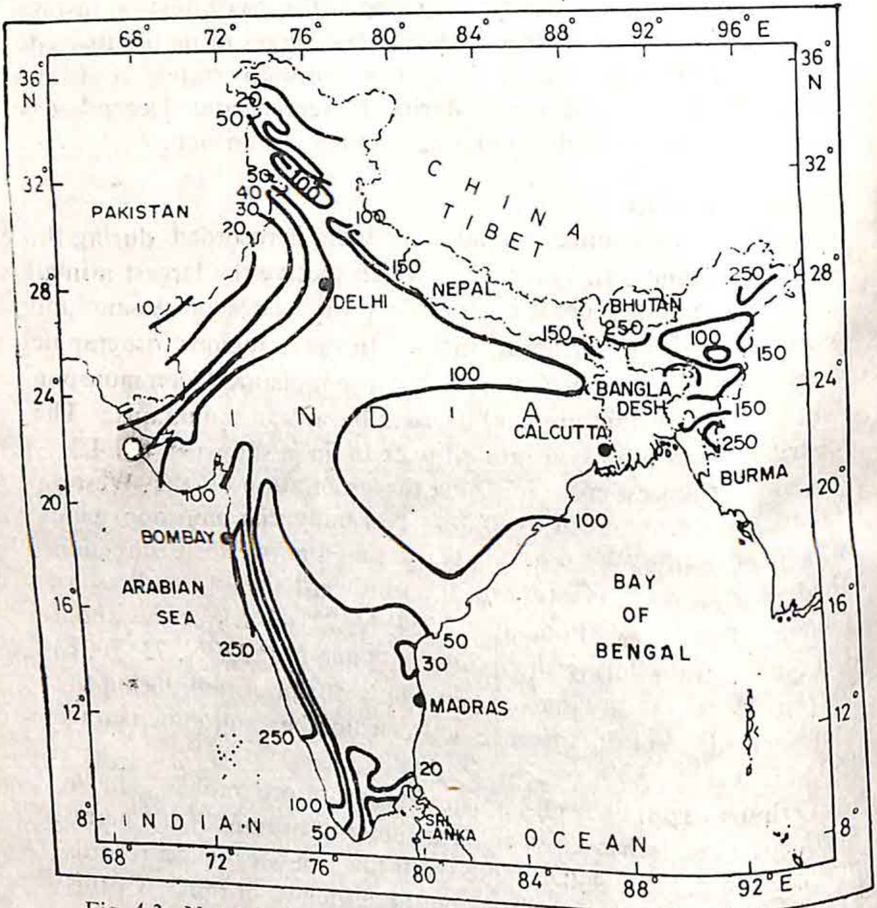

Fig. 4.3—Normal distribution of monsoon rainfall in cm (June-September) over India.

Based upon Survey of India map with the permission of the Surveyor General of India, © Government of India Copyright 1988. The territorial waters of India extend into the sea to a distance of twelve nautical miles measured from the appropriate base line. Responsibility for the correctness of internal details shown on the map rests with the publisher.

is 1141 cm (449 inches). This is a little higher than Cherrapunji's annual rain.

Both Cherrapunji and Mawsynaram are located on the southern slopes of the Khasi hills, the mean height of which is about 1.5 km. There is little doubt that the major part of the

rainfall recorded at these two places can be attributed to orographic features. They are both located at the northern end of a deep valley running from the south to the north. When the monsoon winds blow from the south, they are trapped within the valley and eventually strike Cherrapunji or Mawsynaram in a perpendicular direction at the end of the valley. It is not surprising to find that the heaviest falls occur when the winds blow directly on the Khasi hills.

It is fairly simple to make a rough quantitative assessment of the rainfall, if we assume that the mositure laden winds are bodily lifted. It has been estimated that with an average southerly wind of 40 km per hour (25 mph), a daily rainfall of 447 cm (17.6 inches) could be reasonably expected. This figure is of the right order of magnitude, suggesting that most of Cherrapunji's rain is the consequence of air being lifted as a large body of water vapour.

A curious feature of monsoon rain at Cherrapunji is that most of it falls during the morning hours of the day. It has been suggested that this is partly caused by two different air masses coming together. During the monsoon months the prevailing winds along the Brahmaputra valley generally blow from the east or the northeast. On the other hand, the winds over southern Assam are from the south. The confluence of these two wind systems is usually located in the vicinity of the Khasi hills, but why should this particular feature result in more rain in the morning hours is not well understood. It seems likely that the winds which are trapped in the valleys at night begin their upward ascent only after they are warmed during the day. This explains partially the observed preference for morning rainfall. Apart from orographic features, atmospheric convection plays an important role during the monsoon and the period just preceding it.

In the pre-monsoon months of April and May, for example, parts of northeast India, especially West Bengal, Bihar and Assam experience severe pre-monsoon thunderstorms. They are given the picturesque name of "Nor' Westers" because they appear to come from a northwesterly direction. In Bengal, they are known as a 'Kal Baisakhi', meaning a mass of dark clouds in the month of Baisakh. The rainfall associated with these thunderstorms is of a

transient nature. The intensity of precipitation is high—often as much as 5 cm of rain are recorded in one hour—but the rainfall is of short duration. Monsoon rain is of a different genre. It is in the nature of continuous rain spread over days, and the intensity of precipitation is not as high as that of convective rain. But there are occasions of "cloud bursts" within a spell of monsoon rain.

Cloud bursts lead to distressing situations. A cloud burst over New Delhi, for example, between 20th and 21st July, 1958 provided 28·6 cm (10·5 inches) of rain in a 24-hour period ending at 08·30 hrs. I.S.T. of July 21. About 68% of rainfall in 3 hours was towards the early part of the morning of July 21. The rate of precipitation during this period was over 7·5 cm (3 inches) per hour. This rainfall was about the highest ever recorded over New Delhi in 84 years. Many parts of the city were inundated.

Intense rainfall is often generated by low pressure systems which appear to be embedded within the monsoon trough. These systems are seen as zones of falling pressure with counter-clockwise winds round them. They do not have the intensity or the dimensions of a proper depression, yet the rainfall generated by these shallow systems is high. An example was provided recently by a spell of very heavy rain over the city of Jaipur in east Rajasthan between 18th and 20th July, 1981. Normally, Jaipur is on the verge of the semi-arid tracts of the Thar desert of Rajasthan. Its annual rainfall is only 59·8 cm, but in the 3-day spell between 18th and 20th July, 1981, 82·4 cm of rain were recorded. Thus, the annual rainfall was exceeded in only 3 days. It was a record for Jaipur.

We have confined our attention so far to short period fluctuations of monsoon rainfall. They reflect spells of heavy rain with a duration of three to seven days. Of the longer term changes, a feature of considerable interest concerns the variability of monsoon rainfall. There are parts of India where the pattern of rainfall shows little or no fluctuation from year to year. These are regions of no variability. For such climatic zones one can predict with reasonable certainty that the seasonal rainfall will be only slightly different from its expected normal value. The lowest variability of monsoon rainfall is observed over northeast India. The rainfall in these parts is invariably within 10% of its

normal climatological expectation. This is not an unexpected result because rainfall in this region is caused by orographic features.

On the other hand, rainfall over western India, Rajasthan and Tamil Nadu is highly variable. Interestingly enough, these are the very regions which receive the least amounts of monsoon rain. Hence, one can conclude that wherever the monsoon rainfall is scanty, it is highly variable. This fact is important to those who wish to develop long range prediction techniques. Clearly, if the rainfall is not likely to depart from its climatological norm, there is little to be gained in trying to predict it in advance. In view of this, it is not surprising to find that long range prediction of rainfall has been largely confined to northwest India and to the Indian peninsula; both are regions of high variability.

Of the other macro-scale variations of rainfall, it is observed that monsoon circulation features attain their maximum northward extension around the beginning of August. This is an average of the two dates which correspond to the establishment of the monsoon over the whole of India (July 15) and the beginning of its withdrawal (September 1). It is also observed that around mid-August we usually have the maximum frequency of 'breaks' in monsoon rainfall. In fact, there appears to be considerable difference in the rainfall patterns of July and August. This points to the fact that there are wide day-to-day regional variations of monsoon rainfall. The monsoon cannot be regarded as one of uninterrupted rain for the whole country.

Normal winds and pressure

Upper winds at meteorological stations in India are largely determined with the help of balloons. Notwithstanding its simplicity, the balloon is an important vehicle for tracing the large scale movements of air. A simple rubber balloon, when inflated with hydrogen, can be made to rise about 16 km before it bursts. The higher we go hydrogen, which gives buoyancy to the balloon, begins to expand. By the time a balloon reaches 16 km, the gas expands to about 10 times its initial volume. With the help of special rubber envelopes of great uniformity and strength, it is possible to send up balloons to 40 km, but the cost becomes prohibitive for routine use at meteorological stations. For most

practical purposes our information about normal winds is limited to 16 km.

The path of a balloon is usually tracked at the ground by a theodolite. This raises an additional difficulty because in the monsoon months the sky is obscured by low clouds. The balloons soon become invisible and optical tracking of balloons is well-nigh impossible. To get over this difficulty, radar tracking of balloons is now increasingly used but, because of its high cost, the network of upper winds is not as dense as one would like it to be. Despite these limitations it has been possible to collect a considerable volume of observational data on upper winds over the years.

The distribution of winds over the earth is closely related to the distribution of barometric pressure. There is a relation between the wind and the pressure distribution, which goes by the name of Buys Ballot's Law. According to this law, an observer in the northern hemisphere who stands with his back to the wind will have lower pressure to his left and higher pressure to his right. In the southern hemisphere the opposite situation prevails, that is, an observer has lower pressure to his right and higher pressure to his left. With the help of this law we may infer that, in the northern hemisphere, the circulation round a centre of low barometric pressure (cyclone) will be always counter-clockwise in direction. When we move to the southern hemisphere however, the circulation round a cyclone is clockwise, in agreement with Buys Ballot's Law. In practice it is observed that the surface wind, while blowing in the general sense indicated above, blows slightly across lines of equal pressure (isobars) at an angle of 20° to 30°. This is because of friction exerted by the ground on the overlying air. The frictional force tends to upset the balance between forces generated by the gradient of pressure and the rotation of the earth.

The meteorologist usually measures pressure with a mercury barometer. He deduces the pressure from the height of a column of mercury. In many physical sciences, the pressure is expressed as a length. The standard pressure at sea level, for instance, is 760 mm of mercury. But in meteorology, it is necessary to deal with much wider variations in pressure; consequently, it is convenient to use a different unit called the millibar (mb).

The 'millibar' expresses pressure as a force acting over an area

of unit cross-section. The unit of force in most physical sciences is the dyne. This represents a force which, applied to a mass of 1 gram, produces a unit acceleration of 1 cm/s^2. The unit of pressure should logically be a dyne per square centimetre. But, because this is too small for meteorological purposes, a larger unit, known as the millibar (mb) has been designed. This represents a force of a thousand dynes per square centimetre. It may be shown with the help of a little arithmetic that a standard pressure at sea level of 760 mm is equivalent to 1013.2 millibars (mb).

To obtain a three-dimensional picture of the physical state of the atmosphere, it is customary to present normal wind and pressure data in the form of cross-sections of the atmosphere at different pressures or altitudes. In figures 4.4 and 4.5 we present normal winds at 1000 mb and 300 mb for July, which is a typical monsoon month. These figures are representative of the monsoon wind field at sea level and at about 9 km.

Let us consider the mean pattern of winds at the lower level first. We note that the air, which moves upto the equator from about 10°S from the southeast, changes direction on crossing the equator and approaches the west coast of India from a westerly or southwesterly direction.

This change in direction may be explained by the earth's rotation. It can be shown on theoretical grounds that the effect of the earth's rotation is to generate a force, known as the Coriolis force, which tends to deflect the wind to the right in the northern hemisphere. On crossing the equator then, a large current of air acquires an anticyclonic (clockwise) rotation and approaches the west coast of India from a westerly or southwesterly direction.

If we refer to the normal winds at the higher level (300 mb), we notice that the winds over India have become easterly. There is a complete reversal in the wind direction from westerlies at lower levels to easterlies as we go higher up in the atmosphere. In fact, it will be noticed that at 300 mb most parts of the Indian sub-continent come under the spell of an extensive anticylone which lies over the monsoon wind system at lower levels. The height at which the wind changes direction is between 500 and 400 mb (6 to 7 km).

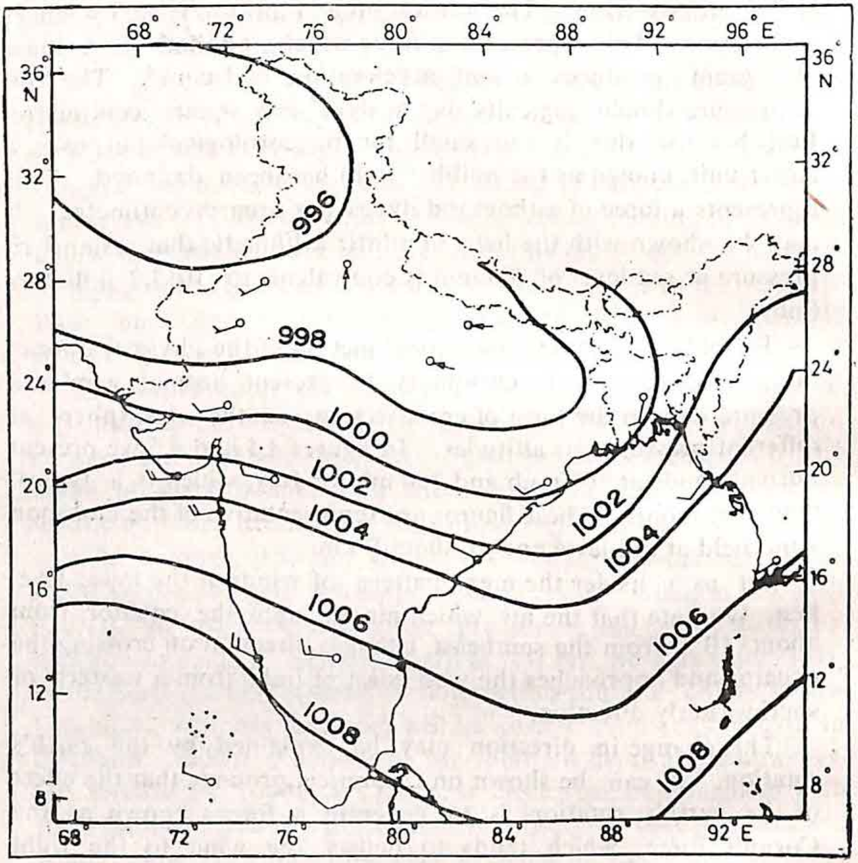

Fig. 4.4—Normal monsoon winds and pressure (mb) at sea-level (1000 mb) in July.

The territorial waters of India extend into the sea to a distance of twelve nautical miles measured from the appropriate base line. Responsibility for the correctness of internal details shown on the map rests with the publisher.

The Easterly Jet Stream

As mentioned briefly in an earlier chapter, the advent of an Easterly Jet Stream and the northward movement of the subtropical westerly jet, is a feature of considerable interest.

The belt of strong easterly winds blow along the southern periphery of an upper tropospheric anticyclone. This narrow belt of strong easterlies is observed between 200 and 100 mb. These

easterly winds, which often record speeds exceeding 100 knots, are known to meteorologists as the Easterly Jet Stream of the tropics. The core of the Easterly Jet Stream is located at about 150 mb (13 km).

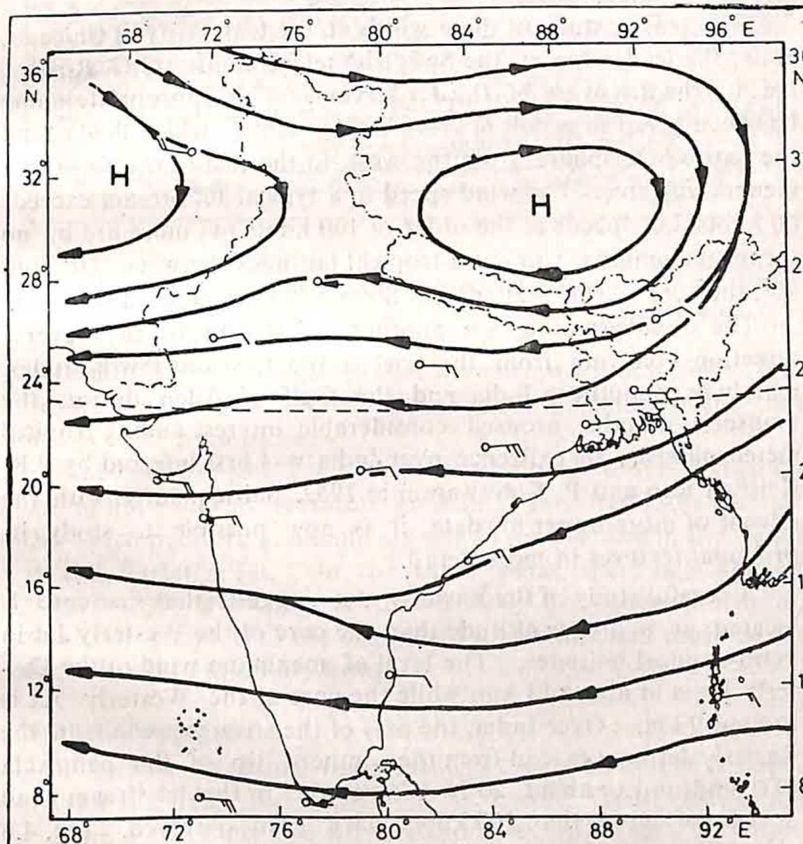

Fig. 4.5—Normal monsoon winds at 300 mb (9 km) for the month of July. The territorial waters of India extend into the sea to a distance of twelve nautical miles measured from the appropriate base line. Responsibility for the correctness of internal details shown on the map rests with the publisher.

Towards the end of the Second World War, rapid developments in the field of aviation provided a much needed impetus for exploring the upper atmosphere. At about this time, U.S. bomber

squadrons flying over Japan often met with strong head winds which were of the same order of magnitude as the speed of their aircraft. Such fast moving air currents had not been encountered before, and aviators were not a little perturbed over their mysterious appearance.

An intensive study of these winds at the University of Chicago, under the leadership of the Swedish meteorologist, C.G. Rossby, led to the discovery of the Jet Stream. This appropriate name has been given to a belt of very strong winds, which flow round the entire hemisphere from the west to the east in the form of a meandering river. The wind speed in a typical Jet Stream exceeds 60 knots, but speeds of the order of 100 knots or more are by no means uncommon. In extra-tropical latitudes between 20° and 40°, the core of the Westerly Jet Stream is located at about 9 km.

The discovery of a yet another jet stream in the reverse direction—blowing from the east to the west—at low latitudes, mainly over southern India and the Gulf of Aden during the monsoon months, aroused considerable interest among tropical meteorologists. Its existence over India was first inferred by P.R. Krishna Rao and P. Koteswaram in 1952. Subsequently, with the advent of more upper air data, it is now possible to study its principal features in more detail.

A careful study of the Easterly Jet suggests that its core is located at a higher altitude than the core of the Westerly Jet in extra-tropical latitudes. The level of maximum wind in the Easterly Jet is at about 13 km, while the core of the Westerly Jet is around 9 km. Over India, the axis of the strongest winds in the Easterly Jet may extend from the southern tip of the peninsula (Trivandrum) to about 20°N (Calcutta). In this jet stream wind speeds of more than 100 knots have been observed. Fig. 4.6 shows the axis of the Easterly Jet at 200 mb (12 km) on a typical monsoon day.

It will be observed from fig. 4.6 that, in addition to an Easterly Jet over peninsular India, there is the Sub-Tropical Westerly Jet Stream to the north of the Himalayas. In winter, it is located along the southern slopes of the Himalayas. But, with the advent of the monsoon, it exhibits a sudden shift to the north. When the monsoon establishes itself over India, its normal position moves far to the north of the Himalayan barrier. The periodic move-

ments of the sub-tropical jet stream often provide useful indications of the onset and subsequent withdrawal of the monsoon.

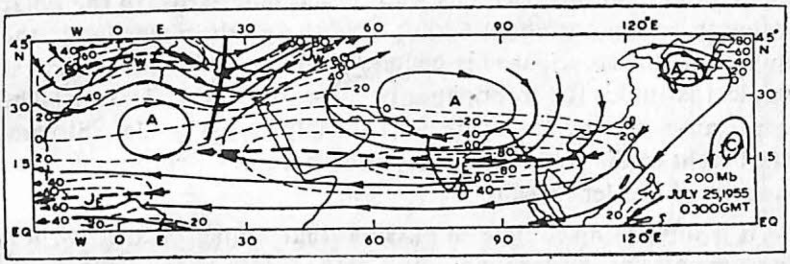

Fig. 4.6--Axis of the Easterly and Westerly Jet Stream at 12 km. (After Koteswaram.) Jet axes are shown by the thick broken lines.

Indeed it has been suggested that the northward movement of the sub-tropical jet provides one with the first indication of the monsoon's onset over India.

The sub-tropical jet stream is observed above a region of strong horizontal temperature gradient. But over India the horizontal temperature gradients are not so pronounced in the vicinity of the Easterly Jet. On the other hand, there is evidence to suggest that the location of the Easterly Jet moves north and south in phase with the northward and southward movement of the axis of the monsoon trough.

Normal Temperatures

The manner in which temperature varies with height is important for the study of weather. It is known from observations that, on an average, temperature decreases with height at approximately 1°C per 100 metre from the ground upto a considerable height. Eventually a stage is reached when the temperature ceases to decrease with height. At greater heights, the temperature remains constant, or shows a slight increase.

It is convenient to divide the atmosphere into two broad thermal zones. The lower region is known as the "troposphere", in which the temperature decreases with height. The upper region, in which the temperature is constant or increases slightly, is called the "stratosphere". The surface of separation between

the troposphere and the stratosphere is known as the "tropopause". The height of the tropopause has an interesting variation with latitude. It is highest at the equator where it is located at about 18 km, but decreases as we proceed polewards. In the polar region it is found at about 6 km above the earth. Over India, the height of the tropopause is generally 16 km, while in the extra-tropical latitudes the tropopause is observed near 11 km. The tropopause often changes its level sharply across a Jet Stream. The height of the tropopause may even become discontinuous near the axis of the Jet Stream.

It is interesting to note in passing that some of the earliest concepts of the atmosphere came rather close to the truth as we know it now. There is evidence to suggest that Aristotle divided the sphere of air into three regions. The first and lowest region was warmed by the sun's rays and the earth's heat. The second region was cold, where the 'watery meteors condensed into clouds'. Lastly, the uppermost region was hot because of its contact with the 'sphere of Fire'.

The modern picture of the atmosphere is certainly more complex than Aristotle's view but, historically speaking, it seems remarkable how close he was to reality with very little visual aid.

The composition of the atmosphere, as we know it today, is shown in fig 4.7. We note that the depth of the stratosphere is of the order of 50 km. Between 50 and 85 km we have another region where the temperature decreases fairly rapidly with height. This is named the Mesosphere, where photo-chemical actions between ionized particles become predominant. Above the mesosphere, we have the Thermosphere where temperatures again begin to rise with height.

In India our knowledge of upper air temperatures has been largely derived from radiosondes. This is an automatic radio-transmitter broadcasting a signal that is coupled to sensors. These sensors are sensitive to different meteorological variables, such as, the pressure, the temperature and humidity. There are many devices by which one can couple a sensor to the transmitting unit of a radiosonde. As the radiosonde is attached to a balloon, its range is also limited to that of the balloon. For most practical purposes this is 16 km.

There are two principal features of the normal temperature

pattern during a monsoon month. Firstly, we find a zone of high temperatures over the semi-arid regions of northwest India. This region of high temperatures is built up gradually during the

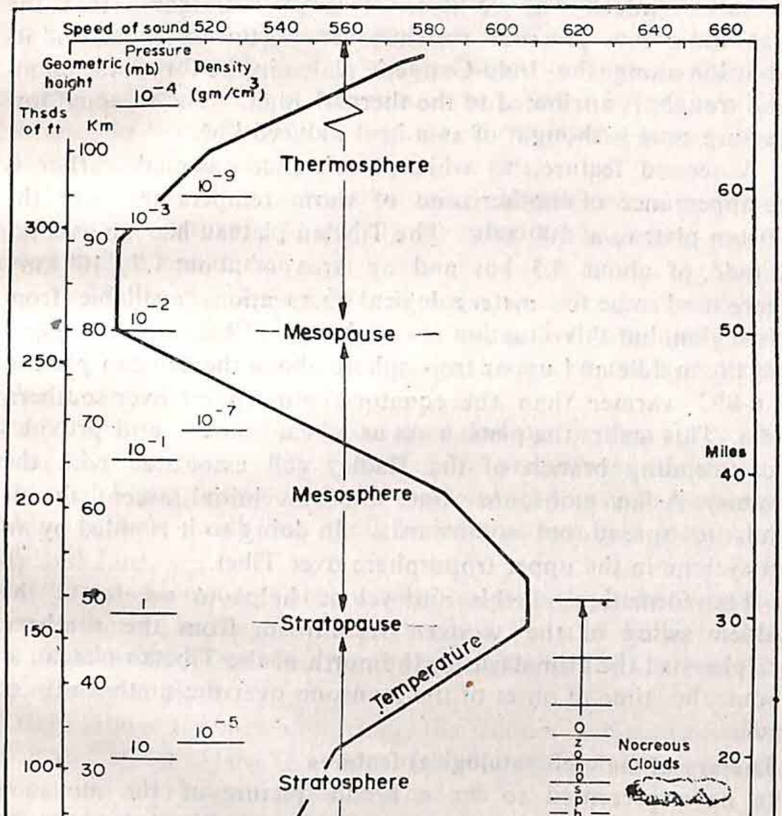

Fig. 4.7—Composition of the Atmosphere.

pre-monsoon months of May and June. The interesting feature is that the geographic location of the thermal high appears to coincide fairly well with the position of an area of low barometric pressure. Indeed, the coincidence is so remarkable that the quasi-static low pressure system over northwest India, and its extension along the Indo-Gangetic plains in the form of a monsoon trough, is attributed to the thermal high. The seasonal low pressure zone is thought of as a heat induced low.

A second feature, to which a reference was made earlier, is the appearance of another zone of warm temperature over the Tibetan plateau at 500 mb. The Tibetan plateau has an average altitude of about 4.5 km and an area of about 1.7×10^6 km^2. There used to be few meteorological observations available from this region, but this situation is now better. Observations suggest that the middle and upper troposphere above the Tibetan plateau is 6-8°C warmer than the equatorial atmosphere over southern India. This makes the plateau act as a heat source, and provides the ascending branch of the Hadley cell associated with the summer Asian monsoon. But, after an initial ascent the air tends to spread out southwards. In doing so it is aided by an anticyclone in the upper troposphere over Tibet.

The formation of this anticyclone helps to accelerate the sudden swing of the western Jet Stream from the southern periphery of the Himalayas to the north of the Tibetan plateau at about the time of onset of the monsoon over the southern tip of India.

Summary of main climatological features

We have presented so far a broad picture of the monsoon circulation as we know it today. It is desirable at this stage to attempt a summary of its salient features for we must first know the facts before we attempt to build a theory to explain the monsoon. What, then, are the main characteristics of the Indian summer monsoon which need an explanation ? We have :

(*i*) The summer monsoon sets in over the extreme south of Indian peninsula on the first of June. The arrival of the monsoon is a gradual process, with a period of transition spread over a week or more.

(*ii*) Subsequently, the monsoon advances along the west coast and into West Bengal and Assam in northeast India. The Bay

of Bengal branch is deflected by the orientation of the mountains into the Indo-Gangetic plains of north India.

(*iii*) The normal duration of the monsoon varies from a hundred to hundred and twenty days. It begins to withdraw from northwest India by mid-September.

(*iv*) Over seventy per cent of India's annual rainfall is recorded in the summer monsoon months. Much of the rainfall is caused by the fortuitous orientation of mountain barriers, but convective phenomena play an important role.

(*v*) The variability of monsoon rain is highest over northwest India and Rajasthan. These are the areas which receive small amounts of monsoon rain.

(*vi*) Monsoon winds approach India from a southwesterly direction. By the time the monsoon has established itself, the depth of the monsoon current is about 6 km off the west coast of India.

(*vii*) Northwest India is an area of low barometric pressure during the monsoon. This region of low pressure coincides with a thermal high (a region of high temperatures), which gradually builds up over northwest India in the pre-monsoon months of May and June.

(*viii*) An extension of the seasonal low into the Indo-Gangetic plains is known as the Monsoon Trough. The axis of the trough shows periodic movements to the north and to the south of the Indo-Gangetic plains.

(*ix*) Above the monsoon winds, the Indian sub-continent is dominated by an extensive anticyclonic circulation. The reversal of the wind field occurs at about 6 km.

(*x*) Along the southern edge of the anticyclone, we encounter a narrow belt of strong winds at about 16 km above sea level. This is well-known as the Easterly Jet Stream of the tropics. Far to the north of the Himalayas is the sub-tropical jet blowing from the west to the east. The axis of this jet is located along the southern slopes of the Himalayas in winter, but suddenly shifts northwards with the advent of the monsoon.

(*xi*) A region of high temperature is usually observed over Tibet. This provides the heat source, and ascent, for the Hadley Cell. Subsequently, the air spreads southwards aided by an upper tropospheric anticyclone over Tibet.

CHAPTER V

THE WINTER MONSOON

THE MAIN CONCERN OF this book is with the summer monsoon of India, but it is helpful to consider the winter monsoon as well for two reasons. Firstly, as we have seen, the winter monsoon corresponds to the night cycle of the sea breeze described in an earlier chapter; the winter monsoon, like its summer counterpart, is an important component of global monsoons. The second point of importance is its capacity to generate seasonal rainfall over the southeastern parts of the Indian peninsula, Sri Lanka, Malaysia and Indonesia. The heaviest rainfall in these countries occurs in the northern winter months of December and January around the time of the winter solstice.

The global features of the Asian summer and winter monsoons are similar in many ways. Intense convection over Indonesia generates precipitation and warming of the atmosphere by the latent heat that is released by the conversion of water from the vapour to the liquid phase. This is then an area of ascent corresponding to the rising limb of the Hadley cell. The ascending air spreads out both to the north and the south. A point of difference between the summer and winter monsoons may be noted here. The ascending air over Tibet largely spreads southwards because its northward passage is blocked by the Himalayas. But, as the ascending branch of the winter Hadley cell is not hindered by mountains, the air is able to spread both to the north and the south.

The northward moving air descends over Siberia and China, a region that is dominated by a large anticyclone. This is interesting because the descending branch of the summer Hadley cell

also occurs over a region of high pressure, namely, the Mascarene High. The southward moving air is associated with a return current that forms the Australian monsoon.

Jet streams in the upper troposphere are features of both the summer and winter monsoons of Asia. The tropical Easterly Jet dominates the summer monsoon, while the subtropical Westerly Jet is a characteristic feature of the winter monsoon.

Walker cells also appear with the winter monsoon. The descending branch of the Walker cell is believed to lie over the eastern Pacific, but this has not been well documented so far.

As with the summer monsoon, the winter monsoon is not one of uninterrupted rain. On the contrary, the rainfall occurs in spells. These spells of heavy rain are accompanied by strong surges of cold air that emanate from a large anticyclone located over Siberia and the adjoining parts of China. Occasionally, the outbreak of cold air from the Siberian anticyclone is associated with the development of a family of low pressure systems over the equatorial regions of the South China Sea and the western Pacific. The disturbances move slowly westwards and bring widespread rain, particularly over Sarawak and the eastern coast of Malaysia.

On some occasions, the low pressure systems intensify into depressions. The rainfall associated with these depressions is heavy and often cause severe floods. A depression of June 3, 1971 was estimated to have caused torrential rains and floods over Malaysia. The total damage was estimated to be about 25 million U.S. dollars.

Meteorologists have often wondered what causes these low pressure systems to intensify. The interaction between low pressure systems and an outbreak of cold air from the Asian anticyclone is not well understood at present. Many scientists have tried to introduce the concept of instability at the interface of two air masses having different meteorological characteristics. The first air mass is an extension of the Asian anticyclone, while the second one is generated by the northern near-equatorial trough.

Measurements of upper winds suggest that on some occasions the necessary conditions for intensifying an incipient disturbance

are indeed satisfied, but we have not been able to show that these conditions are sufficient by themselves.

Over Tamil Nadu in peninsular India, the winter or the northeast monsoon is from October to December.

Tamil Nadu consists of eleven districts, namely, Chingleput, North Arcot, South Arcot, Tanjore, Tiruchirapalli, Madurai, Ramand, Tirunelveli, Salem, Coimbatore and the Nilgiris. It stretches as an extensive hinterland to two of India's major ports, Madras and Cochin.

The entire region is shielded by a mountain range, the Western Ghats, from the rain bearing winds of the summer or southwest monsoon. Consequently, it depends mainly on the northeast monsoon rains for agricultural production.

As with the southwest monsoon, the onset of winter rains over Tamil Nadu is a gradual process beginning with a period of transition. The duration of the transition period is about a week. There is, however, one important difference between the summer and winter monsoons of India.

The onset of the summer (southwest) monsoon is much more well defined. It follows a progressively northward movement which can be discerned with reasonable accuracy on weather charts. The onset of the winter (northeast) monsoon, on the other hand, is not so clearly defined. In fact, on many occasions there is no clear distinction between the withdrawal of the summer monsoon over peninsular India and the onset of the winter monsoon. One tends to merge into the other. We will, therefore, not present charts to depict the normal dates of onset of the northeast monsoon, but proceed on the understanding that the winter months of October to December represent the northeast monsoon.

The average rainfall in Tamil Nadu during the northeast monsoon is about 46.5 cm (18.5 in). This represents about 47.7 per cent of the total annual rainfall. In the following table (Table 5.1) we present a few details of the rainfall of Tamil Nadu taken from a study by Krishna Rao and Jagannathan (1953).

The interesting feature brought out in table 5.1 is that the coastal districts of Tamil Nadu, viz., Chingleput, South Arcot, Tanjore, Ramnad and Tirunelveli receive over 50 per cent of their total annual rainfall from the northeast monsoon. The

Table 5.1
Rainfall of Tamil Nadu

(After Krishna Rao and Jagannathan, 1953)

District	Area in sq. km.	Number of rain-gauge stations*	Normal rainfall (cm)					
			Annual	Oct.	Nov.	Dec.	October to December	
							Actual	% of annual
1. Chingleput	7975	14 (6)	118.62	26.39	30.91	11.84	69.14	58.3
2. South Arcot	13512	17 (6)	119.02	22.99	27.99	14.02	65.00	54.6
3. Tanjore	9609	26 (10)	112.75	20.55	28.73	16.99	66.27	58.8
4. Ramnad	5449	19 (6)	80.54	17.93	18.19	8 81	44.93	55.8
5. Tirunelveli	13957	21 (7)	75.77	16.66	20.04	10.92	47.63	63.2
6. Madurai	17086	19 (5)	81.08	18.31	15.34	5.82	39.47	48.7
7. Tiruchirapalli	9407	20 (3)	88.14	18.16	15.24	7.16	40.56	46.0
8. North Arcot	19129	14 (6)	96.55	16.53	16.00	5.89	38.43	38.8
9. Salem	19503	25 (8)	82.12	15.29	10.21	3.02	28.52	34.7
10. Coimbatore	20202	23 (7)	85.37	17.60	11.48	3.58	32.66	38.3
11. Nilgiri	2481	11 (2)	189.10	25.22	17.58	6.71	49 50	26.2
Tamil Nadu			98.75	19.15	19.10	8.76	47.01	47.7

*Number of rain gauge stations have changed over the years. Figures in parenthesis indicate smallest number of rain-gauges.

remaining districts in the interior of Tamil Nadu receive, on an average, about a third of their annual rainfall during the north-east monsoon.

Storms and depressions

Similar to the situation prevailing over Malaysia and Indonesia, most of the rainfall over peninsular India during the winter monsoon is closely associated with the westward passage of storms and depressions. These vortices are remnants of low pressure systems that affect Malaysia, cross over into the Bay of Bengal and then move westwards over Tamil Nadu. Unfortunately, it is not possible to establish continuity of motion on all occasions because of lack of data. In table 5.2 we provide the monthly frequency of cyclonic storms and depressions in the 70 year period 1891–1960.

Table 5.2

Monthly frequency of Cyclonic Storms and Depressions in the 70 year period 1891—1960

	October	November	December
Bay of Bengal			
Storms	53	56	26
Depressions	56	32	18
Arabian Sea			
Storms	17	21	3
Depressions	11	10	5

It may be observed that the number of storms is more numerous in October and November. The track of these depressions and cyclonic storms appear to lie along a channel bounded by the latitudes of 5° and 15°N. Usually, storms which have a track south of 15°N enter the Arabian Sea from the Bay of Bengal after traversing the Indian peninsula. They have a tendency towards intensification as they emerge into the Arabian Sea. A good number of storms recurve towards the northern sectors of the west coast of India after they enter the Arabian Sea. Unlike the depressions of the summer monsoon, the heaviest rainfall tends to occur on the northern sectors of depressions and cyclonic storms during the winter monsoon.

The dependence of winter rain on the number of cyclonic storms and depressions is very pronounced in the southern parts

of the Indian peninsula. This is more so in the coastal districts than in the interior parts of the peninsula. A map to depict normal winter rainfall over the Indian peninsula is shown in fig. 5.1. Not unexpectedly, the highest seasonal rainfall of around 75 cm between October and December occurs along the southeastern

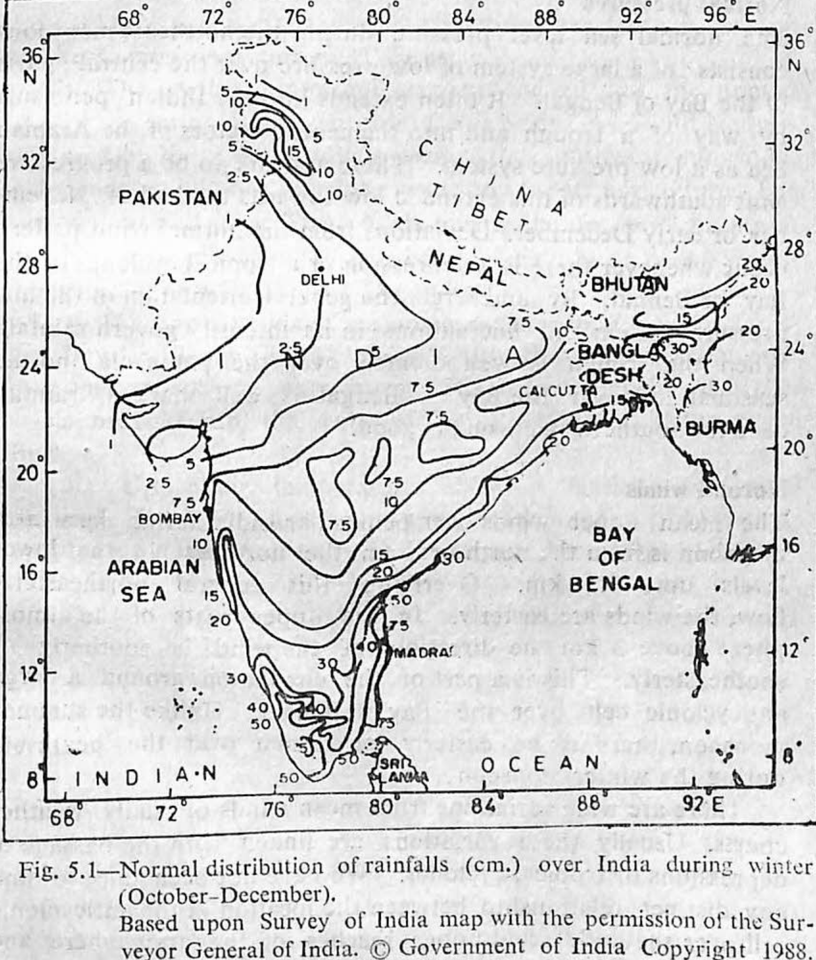

Fig. 5.1—Normal distribution of rainfalls (cm.) over India during winter (October–December).

Based upon Survey of India map with the permission of the Surveyor General of India. © Government of India Copyright 1988. The territorial waters of India extend into the sea to a distance of twelve nautical miles measured from the appropriate base line. Responsibility for the correctness of internal details shown on the map rests with the publisher.

coast of Tamil Nadu and the adjoining parts of south Andhra Pradesh. Thereafter, there is a gradual decrease in rainfall except for a small area towards the north–east of Kerala. This is because of the presence of the Western Ghats which force the normal easterly winds to rise above them.

Normal pressures

The normal sea level pressure during the northeast monsoons consists of a large system of low pressure over the central parts of the Bay of Bengal. It often extends into the Indian peninsula by way of a trough and into the central sectors of the Arabian Sea as a low pressure system. There appears to be a progressive shift southwards of this extended low towards the end of November or early December. Deviations from the normal wind pattern occur whenever there is a depression or a tropical cyclone in the Bay of Bengal. By and large, the general orientation of the low pressure system and fluctuations in its intensity govern rainfall. When the trough is well defined over the peninsula and the seasonal low over the Bay of Bengal is well marked, rainfall over the southern peninsula is good.

Normal winds

The mean upper winds over peninsular India during the winter monsoon is from the north or from the northeast in the lower levels upto 1.5 km. Overlying this general northeasterly flow, the winds are easterly. In the upper parts of the atmosphere above 3 km the direction of the wind is southerly or southeasterly. This is a part of the circulation around a large anticyclonic cell over the Bay of Bengal. Unlike the summer monsoon, there is no easterly jet stream over the peninsula during the winter monsoon.

There are wide variations from mean winds on daily weather charts. Usually these variations are linked with the passage of depressions or tropical cyclones. We have not been able to find any distinct relationship between the location of the anticyclonic cell over the middle and upper reaches of the atmosphere and rainfall over the peninsula.

Normal temperature distribution

The warmest areas in Tamil Nadu during the winter monsoon

are located approximately near 10°N. The temperature then falls off rapidly as we proceed northwards. The northward gradient of temperature increases steadily with height upto 300 mb, at which level the temperature drops by as much as 8°C as we proceed north from 20°N to 30°N.

Summary of main features of the Northeast Monsoon

We may now summarise the main climatological features of the northeast monsoon. In brief, they are :

(*i*) The northeast monsoon accounts for 50% of the annual rainfall in the coastal districts of Tamil Nadu. As Tamil Nadu is sheltered by the Western Ghats during the southwest monsoon, it depends mainly on the winter monsoon for its agriculture. The average rainfall over Tamil Nadu during the northeast monsoon is of the order of 47 cm (18.5 inches). The duration of the winter monsoon is from October to December.

(*ii*) The upper winds of the northeast monsoon sweep across the Bay of Bengal in a clockwise direction. The circulation is round an anticyclone centred near 25°N at lower levels. At higher levels near 500 mb, tha centre of the anticyclone shifts to central Burma.

(*iii*) Upper air temperatures show a marked northward gradient. The warmest temperatures are found near 10°N but, as we proceed northwards to 30°N, the temperature drops by 8°C.

CHAPTER VI

THE PHYSICS OF MONSOON RAIN

A GOOD WAY TO begin the chapter might be to ask ourselves : What causes rain ? We know that one of the major constituents of the atmosphere is water. But, water is found in several different forms. We see it coming down as liquid drops of rain, or in a frozen state in the form of snow crystals. It also exists in a gaseous form as water vapour. To this we should add a fourth phase, namely, water in a super cooled state. In this phase, water retains its liquid form even at temperatures well below its normal freezing point (0°C).

The existence of water in four different states adds not a little to our difficulty in understanding physical processes. It has been estimated that during a typical monsoon day about 75,000 million tonnes of water vapour are transported across the west coast of India. If we consider that the plains of India cover an area of 1.5 million square kilometres, then, an average rainfall of about 1.7 cm (0.7 inch) per day over this area would imply that, on an approximate basis, 25,000 million tonnes of water vapour are converted every day into rain. On an approximate basis then, about a third of the water vapour that enters the western coast of India is converted into water—in the liquid phase—which can be used for agriculture and other enterprises for the benefit of the Indian farmer. Clearly, if we are able to hasten, or increase, the efficiency of conversion from invisible vapour to visible liquid the material benefits would be substantial. Our understanding of the physics of rainfall is centred towards an attack on the problem of getting more water through natural processes.

What causes invisible vapour to change into liquid water ? With the help of laboratory experiments it is possible to demonstrate that when water vapour in the air reaches a saturation point, a very large number of liquid drops are produced. If the saturation point is reached at temperatures below the freezing point of water, the vapour changes directly into crystals of solid ice rather than water drops, but this mode of conversion from vapour to solid is comparatively rare in the atmosphere.

The saturation point of water vapour is one of the important parameters that determine the rate at which vapour is converted into liquid water. The question may well be asked : Why should there be a critical point at which vapour changes to water (or ice) ? The answer lies in the fact that, at any specified temperature and pressure, there is an upper limit to the amount of water vapour which a parcel of air can hold. This is not unexpected because, after all, water vapour is only a collection of water molecules in gaseous form. Consequently, if we confine the space available to a parcel of air by specifying its temperature and pressure, it follows that there will be only room for a certain number of water molecules and no more. If we try and force more water vapour into the atmosphere than it can possibly hold, it gets rid of the surplus vapour by making it condense in the form of liquid drops.

We may readily decrease the vapour retaining capacity of the atmosphere by lowering its temperature. Most readers would have noticed the condensation on a tumbler of water when we lower the temperature of the surrounding air by dropping a few pieces of ice into the tumbler.

In the atmosphere the condensation of water vapour gives rise to clouds. A convenient way of describing a cloud would be to picture it as a population of liquid water drops held in animated suspension. Indeed the size of water droplets often determines the shape and growth of an individual cloud.

Condensation Nuclei

Condensation nuclei have an important role to play in the conversion of vapour to liquid water. These are the aerosols which are found floating in the atmosphere. They act as embryos on which the first drops of water form. It is possible to demonstrate, on

theoretical grounds, that the presence of nuclei is necessary for condensation.

We could illustrate this by considering a flat sheet of water; for example, an open trough of water which is exposed to the atmosphere. It is but natural to assume that in such a situation there would be an exchange of molecules across the air-water interface. The molecules of water are constantly moving about. A small proportion of molecules, which move more rapidly than others, escape from the liquid by overcoming the internal molecular forces which bind them together. The situation is much the same as that of a missile which, given sufficient speed, escapes from the earth's gravitational attraction. In a similar fashion, a number of molecules of vapour penetrate the water surface and are captured by the liquid. Eventually a state of equilibrium is set up when the number of molecules escaping from the liquid equals the number of vapour molecules entering it. When this happens the shape of the interface is stable and the vapour pressure acquires a steady value.

Let us consider now a spherical water drop surrounded by an environment of vapour. In this situation the shape of the water vapour interface is curved instead of being flat; consequently, it has a larger area than a flat interface. In view of its larger area, a larger number of molecules are transported across it when a state of equilibrium is reached. Another way of stating this would be to assert that the vapour pressure required to maintain a spherical drop in equilibrium is larger than the saturation vapour pressure over a flat surface. In other words, a certain amount of supersaturation is necessary to maintain a spherical drop in equilibrium.

This interesting observation is due to Lord Kelvin, who derived a mathematical relation between the excess of vapour pressure and the curvature of the interface. From Lord Kelvin's results, we are able to reason that if water vapour was directly converted to liquid droplets—of the size of water molecules—an exceedingly high degree of supersaturation would be necessary. The excess vapour pressures required to maintain such minute droplets in equilibrium would be quite unrealistic. On the other hand, if condensation occurred on a nucleus, the water-air interface would have a much larger radius of curvature. Consequently,

a much smaller degree of supersaturation would be sufficient to initiate the formation of liquid drops.

From this reasoning we see that the presence of nuclei is essential for the process of condensation. As the sizes of nuclei are much larger than the dimensions of a water molecule, a drop once formed by condensation on a nucleus requires a considerably lesser degree of supersaturation. There is, therefore, a reasonably good chance for the drop to survive in the atmosphere. On the other hand, a drop which does not form on a nucleus would need an extremely high degree of supersaturation. It would almost certainly evaporate as soon as it was formed.

It is worthwhile to recall that this fact was well demonstrated in a laboratory experiment by C.T.R. Wilson, the scientist to whose ingenuity we owe the Expansion Cloud Chamber. Wilson was able to show that if we could purify the air sufficiently by removing from it all traces of nuclei, then it was possible to prevent condensation even with very high degrees of supersaturation. Despite relative humidities of the order of 400 per cent, little or no condensation was observed in the absence of nuclei.

It is important to note that the curvature of a liquid vapour interface is one of several factors which detremine the saturation vapour pressure needed to maintain a droplet in equilibrium. It is possible to show, for example, that the degree of supersaturation that is required is considerably smaller for nuclei that are hygroscopic. Hygroscopic substances are those that absorb moisture. In India, hygroscopic nuclei are present in plenty in the form of suspended salt particles or minute drops of acid contained in gases from industrial factories.

When hygroscopic nuclei are present in the air, it is possible for clouds to form with little or no supersaturation. There is evidence to indicate that condensation has taken place even when the relative humidity of the air was below 100 per cent.

A number of experiments have been carried out in recent years to measure the size and concentration of condensation nuclei. These experiments reveal that even on days of fairly good visibility, the number of nuclei in the atmosphere is exceptionally large. The concentration of small condensation nuclei is of the order of a million per litre of air.

The diameter of an air molecule is about a ten millionth part

of a millimetre. The average diameter of a small condensation nucleus is about thousand times larger than an air molecule, that is, about ten thousandth part of a millimetre. In dealing with such small sizes physicists usually use a unit called the micron. A micron is a millionth part of a metre (10^{-6} m). On this scale, the dimensions of different particles, are shown in table 6.1.

Table 6.1
The sizes of Aerosols in the atmosphere

Particle	Diameter (microns)	Mass (grammes)	Number in 1 cu. metre of air
Air molecule	10^{-4}	10^{-22}	10^{25}
Small nuclei	10^{-1}	10^{-15}	10^{9}
Giant nuclei	10	10^{-9}	10^{3}
Cloud droplets	10 to 100	10^{-9} to 10^{-6}	10^{6} to 10^{9}
Rain drops	1000	10^{-3}	10^{3}

The problem of sampling and measuring the diameters of such small particles is a difficult one in experimental physics. We have to design sensitive measuring devices. The instruments must be capable of withstanding considerable stress, because measurements are made from an aircraft. Progress in this field has been largely achieved by instruments which draw in air at a controlled rate through the fuselage of an aircraft. The air stream is then permitted to strike a series of slides arranged in the form of a cascade. The impact of cloud droplets on these slides, which are coated with suitable chemicals, leaves a mark which may be measured under a microscope. This is the principle of a cascade impactor which is used to measure the size of aerosols.

The origin of hygroscopic nuclei in the atmosphere is an intriguing problem. From the little experimental evidence that we have, it appears that such nuclei are largely particles of sea salt. Considering the fact that the earth is largely ocean-covered, it seems natural to assume that most hygroscopic nuclei have an oceanic origin. Meteorologists have believed that very large numbers of salt particles are injected into the atmosphere by the breaking of waves. When waves strike the coastal regions of a large land mass they release a large volume of spray. Millions of minute salt particles are injected into the free atmosphere in this

manner, and they provide nature with a store-house of condensation nuclei. But can we be sure that an adequate number of nuclei are indeed generated by this process ?

Many years ago Sir George Simpson argued that the average rainfall of the earth was about 100 cm per year. This represents 100 cm^3 of water over each square centimetre of the earth's surface. Let us assume that this volume of water is obtained by a combination of cloud drops, each having a radius of 10 microns, and let each drop contain a salt particle as its nucleus. With a little arithmetic it is possible to show that the number of nuclei returned to earth should be of the order of 2.5×10^{10} per year over each square centimetre. This works out to be about 1000 nuclei per second over each square centimetre. The rate at which nuclei are generated by breaking waves should be at least of this order. It was felt that such a high rate of nuclei production was impossible. But, more recently, laboratory experiments by Professor B.J. Mason in England suggest that the bursting of minute bubbles from the sea surface might well produce 1000 nuclei/cm^2 per second, which is the figure required.

The mechanics of rain formation

We have now seen that when water evaporates from the earth's surface, it is carried into the atmosphere as invisible vapour. Mainly because the vapour laden air cools as it rises, a point is reached when the air is no longer able to hold water in the form of vapour. The altitude at which this happens is known to meteorologists as the Lifting Condensation Level. Further cooling by ascent above this level leads to condensation of water vapour in the form of visible clouds. Condensation inevitably takes place on minute nuclei. Eventually, when the cloud droplets become large enough they fall as drizzle or rain.

But, let us consider the stage between the first condensation of vapour and the subsequent formation of rain. We refer to the myriads of floating droplets which we recognise as clouds. There are good theoretical reasons to explain why condensation at temperatures below the freezing point should be in the form of ice crystals. The saturation vapour pressure needed to maintain an ice crystal in equilibrium is considerably smaller than the corresponding value for a water droplet.

Imagine, then, that in the upper reaches of a cloud we have a mixed population of water droplets and ice crystals. We cannot entirely rule out the possibility of finding water droplets at sub-freezing temperatures because, as we can see, water may exist in a super-cooled state. But, because water vapour condenses more readily on an ice crystal at these low temperatures, it follows that the number of water drops will gradually dwindle. Eventually, in the course of time, the entire population should become one of ice crystals. Calculations show that the transformation of a supercooled cloud could be indeed accomplished in a matter of minutes. Moreover, it would not be unreasonable to assume that some of the ice crystals would eventually acquire larger dimensions at the expense of the smaller ones. In the course of time, these larger crystals would fall more rapidly into the warmer lower atmosphere. In these warmer regions we may expect them to melt into rain drops. In the last stage of the cycle, we should see them as raindrops falling from a cloud.

This was the basis of an important theory on rain formation. It was put forward by a Swedish meteorologist, Bergeron in 1933 and was strongly supported by Findeisen in 1938. In substance, their theory asserts that all rainfall was, in effect, melted ice or snow flakes.

The theory had a few attractive features which were well supported by observations. Firstly, observations tell us that rain bearing clouds are of large vertical growth. They are thick enough to reach the higher levels with sub-freezing temperatures, where the ice phase is almost invariably present. Such clouds have been known to exhibit a sudden transformation when rain begins to fall. The upper 'anvil' in a rain bearing cloud suddenly puts on a fibrous appearance, probably because of the presence of a large number of ice crystals, and the freezing process does appear to play an important role in the rainfall observed from such clouds.

But, the difficulty with the Bergeron-Findeisen theory lay in the fact that rain was also observed from clouds which did not extend upto the freezing level. In India, during the southwest monsoon months, quite a few observations have been made of rain from low clouds, whose tops were far below the freezing levels.

In the last decade or so, it was increasingly realised that although the ice phase was important, it was not essential for rain formation. In "warm" clouds, that is, in clouds which did not reach the level of the freezing point, the formation of rain was brought about by coalescence between falling water drops.

Consider a "warm" cloud in which we have a population of cloud droplets, and the ice phase is entirely absent. If we leave out of consideration, for the moment, the random irregular motion of drops by turbulent fluctuations of wind, it follows that all the drops will gradually begin to settle down under the force of gravity. Eventually, each liquid drop acquires a steady speed of fall which is known as its terminal velocity. This represents a state of equilibrium in which the weight of the liquid drop is exactly balanced by resistance of the air. It is of some interest to note that the resistance offered by air follows an interesting law.

For a large and heavy body, such as a brick or a stone, air resistance is clearly of little consequence. But, for extremely light bodies, such as, a feather or a falling drop, air drag is important because it determines the speed of fall. For small water drops of radius less than 25 microns, the terminal velocity is directly proportional to the square of the radius. This important result was predicted by Sir George Stokes more than a century ago. If we consider air at 0°C and at a pressure of 900 mb, Stokes Law gives us a simple relation between the terminal velocity of a drop (V) and its radius (R). This is

$$V = 1.26 \times 10^6 R^2$$

With the help of this equation, we readily see that the fall speed of a cloud droplet of radius 25 microns is about 7.9 cm sec^{-1}. At this rate of fall, a droplet of this size would take about 4 hours to fall through a cloud which is 1000 metres thick.

Stokes Law is not valid for the larger water drops of drizzle or rain. There is as yet no uniform law by which one can relate the terminal velocity of a spherical body with its size. As an example, let us consider an average rain drop whose radius is 0.2 cm. If it obeyed Stokes Law its terminal velocity would be 500 m sec^{-1} but, in reality, its fall speed is only of the order of 9 m sec^{-1}. We have good reason to be not unhappy about this state of affairs. If larger water drops, or small projectiles, such

as hailstones, followed Stokes Law the damage to mankind from falling hail or snow would have been much greater.

Reverting back to the population of floating drops, we can now see that the larger drops would fall at a faster rate because of their greater fall speed. On their downward path, they would encounter several smaller ones which, owing to their smaller fall speed, do not have time to get out of the way. Would these smaller drops, then, be captured by the larger ones ?

A part of the smaller drops would be captured and, in all probability, would combine with the large drop, but there are other mechanisms which are difficult to estimate. In the first place, it is not obvious that two water drops coming together would coalesce. It is difficult to find out, by laboratory experiments alone, what happens when two small water drops meet. The little experimental evidence that there is at present suggests the possibility of a bounce-off following collision. When two drops collide, their respective spherical shapes are distorted. This, it may be shown, involves a gain in the total energy of the system which has to be supplied from an external source. It seems likely that if the difference in the fall speeds of two colliding drops is sufficiently large, the increase in energy could well be provided by the loss in the kinetic energy of the large drop following its impact with a smaller one. Collision would then be followed by coalescence. But, on the other hand, if the colliding drops are of almost equal size, there is no certainty that collision would lead to coalescence.

A considerable volume of theoretical work has been directed to ascertain the rate of rain formation by coalescence. Most theoreticians base their computations on the assumption that each collision between two drops leads to coalescence. The indications provided by these theoretical studies are that, in the initial stage, the growth of a droplet by progressive condensation of water vapour proceeds more rapidly than the coalescence mechanism, until the drop acquires a radius of about 15 microns. Thereafter, the rate of growth by coalescence begins to predominate. It is likely that in clouds of large vertical extent, growth by condensation of water vapour on ice crystals initiates the precipitation mechanism. The subsequent development of rain must be the result of a combination between coalescence and progressive condensation.

Monsoon clouds

Meteorologists classify clouds by their structure. The first classification of clouds is due to Luke Howard who, in 1803, introduced the Latin names Cirrus, Cumulus, Nimbus and Stratus for different clouds. The *International Cloud Atlas* published by the World Meteorological Organization recognises ten main types. Their main features are :

(*i*) *High clouds*

Cloud base at 6 km (20,000 feet) or higher

(1) Cirrus	:	Detached fibrous clouds in the form of white feathers or narrow bands.
(2) Cirrocumulus	:	Thin white layers of high clouds, without shading.
(3) Cirrostratus	:	Transparent white clouds through which halos are often seen.

(*ii*) *Medium clouds*

Cloud base at or above 2 km (7,000 feet)

(4) Altocumulus	:	White or grey layer cloud; sometimes seen in the form of rolls or round globules.
(5) Altostratus	:	Greyish cloud sheet. Halos cannot be seen through Altostratus clouds.

(*iii*) *Low clouds*

Clouds extending from the surface to 2 km

(6) Stratus	:	Generally a grey cloud layer of uniform base.
(7) Stratocumulus	:	Grey or white patches; often appear as rolls or rounded masses of clouds.

To this list we should add

(8) Cumulus	:	Detached clouds with sharp outlines. Rising towers or domes are often seen within a Cumulus cloud.
(9) Cumulonimbus	:	Heavy and dense shower clouds with tops spread out in the form of an anvil.

(10) Nimbostratus : Grey or dark cloud layers from which we observe continuous rain.

An aviator is mainly concerned with turbulence. Moderate or heavy turbulence is observed near Cumulus and Cumulonimbus clouds. These are clouds of strong vertical growth. A survey of observations recorded by commercial aviators in India suggests that the top of Cumulonimbus clouds could be as high as 16.7 km (55,000 feet) during the monsoon. The survey also indicates that more than half (57%) of all instances of turbulence are encountered in the monsoon months. And, on most occasions moderate or heavy turbulence was in the vicinity of Cumulus or Cumulonimbus clouds.

Conditional instability

Cumulus and Cumulonimbus clouds are also associated with thundershowers during the pre-monsoon and monsoon months. These are shower clouds in which electrical discharges can be seen as lightning, or heard as thunder. Thunderstorms develop when large amounts of cloud water, in the form of liquid or solid, are carried upwards to heights where the temperature is frequently less than—20°C.

There is a tendency for thunderstorms to form in lines. These are known as squall lines, and they occur most frequently in the pre-monsoon months of April and May over northeast India. Sometimes multiple squall lines have been observed. Squall lines are the source regions for a series of thunderstorms of considerable violence. The time of maximum occurrence is usually in the late afternoon, or shortly after the time of maximum temperature, but there are also occasions when thundershowers occur in the late night or early morning.

There is no general agreement on the mechanism by which thunderstorms are generated. Thunderstorms represent regions of strong convective motion in the atmosphere. Meteorologists have tried for many years to find some criterion by which one could predict the onset of convection in the atmosphere. The instability of the atmosphere is relevant here.

Under normal circumstances, the density of the atmosphere decreases with height. Consequently, if an element of air is given a vertical displacement upwards, it will have a density

greater than its new surroundings. It will, therefore, tend to sink back to its original position. Thus it follows that if the density of air decreases with height, the atmosphere tends to suppress small vertical displacements of air. In such circumstances, the atmosphere is stable.

But there are occasions when we observe warmer air near the earth's surface and colder air aloft. The vertical variation of air density is then unstable to small disturbances; for in this case an element of air displaced upwards will have a density less than its new surroundings. Consequently, it will be accelerated upwards by its buoyancy.

In the atmosphere pressure decreases with height, so that the upward displacement of an air parcel results in its expansion and a decrease in its density. The atmosphere will be unstable only if the density of the displaced element is less than the density of the ambient air. The critical condition when this happens can be expressed by the vertical gradient of temperature with height. As mentioned earlier, if the lapse rate of the atmosphere exceeds 1°C per 100 metres instability is likely to set in.

The vertical motions arising from instability produce Cumulus or Cumulonimbus clouds. For cloudy or saturated air, the critical lapse rate is less than 1°C/100m, because as the air ascends there is progressive condensation from vapour to liquid water. The transformation in phase from vapour to liquid is accompanied by the release of an appreciable amount of heat, which is the latent heat of condensation. As the release of latent heat adds to the buoyancy of a rising element of air, the critical lapse rate for saturated air is approximately half the corresponding value for dry air, i.e., 0.5°C per 100 metres.

Indeed it may happen that an atmosphere is stable for the ascent of unsaturated air, but is unstable for saturated air. When this situation prevails the atmosphere is said to be conditionally stable. A conditionally unstable atmosphere is interesting from the viewpoint of scale. Imagine an atmosphere that is conditionally unstable being disturbed by a travelling wave or a low pressure system. In such a situation, the atmosphere—in principle—will have to support both the growth of Cumulus clouds and the travelling disturbance. Which will grow faster, the ensemble of Cumulus clouds or the travelling disturbance ? This question

is relevant when we consider monsoon clouds embedded in a depression in the Bay of Bengal for example, or the clouds over Indonesia during the winter monsoon. Theory suggests that in such situations, Cumulus clouds will always grow faster because of their smaller dimension. But, this need not mean that the entire moisture content of the atmosphere will be spent in supporting Cumulus clouds at the expense of a travelling depression because, if that was the case, the depression would soon decay because of friction exerted by the earth's surface. The fact that this does not happen is because there need be no competition between Cumulus clouds and a depression for the available moisture in the atmosphere. Indeed, recent experiments with mathematical models have brought to light the fact that Cumulus clouds support and maintain a depression by replenishing the energy of the latter through the release of latent heat. The latter in turn support Cumulus clouds by providing them with moisture and rising motion. The process by which this happens is known as Conditional Instability of the Second Kind (CISK). It is different from the conditional instability of moist air.

Cumulus clouds are now recognised as important conveyors of heat and momentum from the lower to the upper troposphere in a conditionally unstable atmosphere, especially in the tropics. This fact is of considerable relevance for the construction of mathematical models for weather prediction in the tropics It is usually referred to as the parameterization of Cumulus clouds. What parameterization seeks to achieve is to quantify the vertical transfer of momentum and heat by a Cumulus ensemble. The technical details are beyond the scope of a book of this nature, but it is important to note that two difficulties arise when we try to parameterize Cumulus clouds. First, we have to assume that the clouds have a simple structure. They are assumed to have a common base altitude which is the lifting condensation level, for example. It is also assumed that they entrain, that is, draw in air from the surrounding atmosphere at a specified altitude. The second difficulty is that the time-scale of the evolution of a Cumulus ensemble is assumed to be small compared to the life cycle of a monsoon depression. Many micro-scale effects are consequently eliminated. Despite such limitations, it has been now demonstrated that prediction models for the tropics, and this

includes the monsoon lands, are unlikely to be realistic unless Cumulus clouds are realistically simulated.

Artificial stimulation of rainfall

Towards the beginning of this chapter we mentioned that the atmosphere does not convert gaseous vapour into liquid water in a very efficient manner. Substantial material benefits could be derived if the efficiency could be enhanced.

With this objective in mind, a number of experiments have been designed in the last two decades to induce clouds to release precipitation. Most experimenters have worked on the hypothesis that, if the number of natural nuclei in a cloud could be increased this, in turn, would augment the rate of drop formation and, ultimately, a larger quantum of water would be released from the cloud.

More than twenty years ago, Vincent Schaefer, an American scientist, discovered that a tiny piece of dry ice, when dropped into a cold chamber filled with supercooled cloud, resulted in the formation of a very large number of ice crystals. Schaefer was closely associated in this work with Irving Langmuir, who had earlier won a Nobel Prize for his research in physical chemistry, and they estimated that a pellet of dry ice, about the size of a pea, could produce about 10^{16} ice crystals in its passage through a supercooled cloud.

The first field trials on cloud seeding appear to have been conducted by Schaefer and Langmuir towards the end of 1946. Observations from the ground suggested that snow fell from the seeded cloud for about 0.6 km before evaporating in the dry air. At about the same time, experiments by Kraus and Squires in Australia reported that six out of eight seeded clouds gave radar echoes after seeding, and in four cases rainfall was reported at the ground.

In later years Vonnegut, another associate of Professor Langmuir, found that silver iodide was a more powerful agent than dry ice for producing ice nuclei. Enormous numbers of nuclei could be produced by vaporizing an acetone solution of silver iodide by a burner. There was the possibility that a large number of nuclei could be disbursed from ground-based burners, and a few experiments in this direction appear to have indicated encouraging results.

For warm clouds, whose tops do not extend to the freezing level, attempts have been made to release showers by spraying Cumulus clouds with water or aqueous salt solutions. The objective is to introduce large drops of water or a suitable chemical solution at the base of the cloud to initiate and expedite the coalescence mechanism.

In India, field experiments along these lines were initiated by Professor S.K. Banerji nearly thirty years ago. A series of 35 experiments were made by Professor Banerji in which he reported rainfall varying from drizzle to heavy rain on 28 occasions.

The initial work of Banerji was later extended by A.K. Roy and his collaborators, who fed hygroscopic nuclei into the base of a cloud from ground based sprayers. They used salt solution in their experiments, which were conducted over Delhi. Although there is some doubt over their method of assessment, it was found that their results had a few positive aspects.

These experiments have been continued by Bh. V. Ramana Murty and his collaborators in India over the last decade. On many occasions, there was evidence to suggest a positive effect, namely, an increase in the intensity of radar echoes, and precipitation was observed from aircraft. These experiments were conducted over the Rihand catchment in Uttar Pradesh and the Linganmakki catchment in Karnataka. Experiments were also conducted in Gujarat in the districts of Panchmahal and Sabarkanta during the years 1975-77.

The results have not been statistically significant, because one is not certain whether seeding of clouds conducted over a period of time would provide an increase in rainfall that was, without ambiguity, greater than the natural variability of rainfall in the region.

In view of these uncertainties opinion among many scientists is sharply divided on the efficiency of this technique. The World Meteorological Organization, with its headquarters in Geneva, decided in 1976 to launch a Precipitation Enhancement Project in Spain.

The objective of this international experiment was to provide reliable and authentic information about the possibility of increasing rainfall by artificial seeding of clouds. It was realised that if

we are to demonstrate the success of this endeavour on a statistical basis, these experiments would have to be conducted for several years. This would enable scientists to study whether the increase or decrease in rainfall was an event of chance, or whether it was brought about by the intervention of scientists. The broad objective was to select two sites, one over which seeding operations are conducted, and another site to act as a control location over which no seeding would be done. By this means one could find out if the seeding mechanism was really efficient. This would be the case if, for example, there was substantial rainfall over the seeded area, but little or no rainfall over the unseeded control area.

Difficulties in reaching a firm conclusion on statistical grounds are indeed so great that the operational phase of this experiment is still to commence. Much preparatory work has been done but one is still not certain whether the design of the experiment is the best that one could hope for. Preliminary work does suggest a high correlation between the vertical depth of convective clouds and the rainfall derived from them. But the statistics of this problem might well require multiple control areas to assess chance trends in surrounding areas.

Research on this subject is still in progress. It would be unwise at this juncture to hazard premature inferences.

CHAPTER VII

THE MONSOON EXPERIMENT (MONEX)

IT IS A SURPRISING fact that although the monsoon arrives every year to keep its date with India, its origin is shrouded in mystery. The roots of uncertainty lie in the fact that the equatorial regions, where the monsoon is believed to have its origin, are largely unknown for the meteorology of large-scale air movements. There are not enough facts to build a satisfactory theory.

In view of these difficulties, a number of experiments were the mounted in the last two decades to ascertain the correct facts and dimensions of the problem. In the early sixties, an International Indian Ocean Expedition (IIOE) was launched. The meteorological programme of this experiment was supervised by a centre in Bombay during 1963. In the first two years of its existence, a number of reconnaissance flights were made by aircraft specially fitted with research instruments. One of the flights reconnoitered the wind field around a tropical cyclone off the west coast of India. A feature of this experiment was that it enabled meteorologists to prepare, for the first time, a detailed atlas of upper winds over the Indian Ocean.

Aircraft observations revealed that the depth of the monsoon current was fairly shallow west of the meridian 65°E. It was only 1.5 km deep westward of 65°E. Thereafter, as we proceed eastwards, the depth of the monsoon air rose sharply to about 6 km near the Indian coastline. A suggestion was made, therefore, that there was a discontinuity in the physical characteristics of the

monsoon air along 65°E. In particular, the existence of a temperature inversion between these two sectors of the Arabian Sea was an important observation. Inversions represent a zone of transition between two air masses. It is observed as an abrupt decrease in the lapse rate of temperature. Physically, it implies the existence of warmer air lying over colder air. We refer to such layers as an inversion in the lapse rate of temperature.

The view was expressed that a rapid increase in the depth of the monsoon eastwards of 65°E was the consequence of an orographic barrier along the west coast of India. On approaching the west coast from a westerly or southwesterly direction, the monsoon air was forced to ascend over the Western Ghats. This is a barrier running along the entire coast in a north-south direction. Consequently, the orientation of the barrier forces the moist and comparatively shallow monsoon air to rise abruptly to a height of around 6 km. It is still not clear how far westwards can we extend the influence of this physical barrier. The question may be asked : would the air at 65°E still feel the influence of the Western Ghats ? This is an unresolved problem.

Another view that was put forward with some strength suggests that the inversion in temperature west of 65°E was really the result of subsiding air motion. It is a commonly observed fact that the systematic ascent or descent of air over large areas is associated with significant changes in the lapse rate of temperature. There is a fairly good reason to explain why this should be so. When we observe large scale descent of air, a column of air tends to shrink vertically because the air descends from a region of low pressure to one of high pressure. The contraction in the vertical column of air is accompanied by warming, or a decrease in the lapse rate of temperature. On certain occasions of marked descent, the decrease in lapse rate may even become an inversion in the vertical gradient of temperature. Consequently, if the observed inversion along 65°E was the result of subsiding air, it would lead us to infer that the semi-arid regions of West Pakistan and Arabia were indeed zones of large scale descending motion.

Interesting questions of this nature have led meteorologists to realise that problems of the monsoon could not be answered unless more field experiments were conducted over the oceanic regions.

This realisation led to two further experiments in 1973 and 1977. These experiments were conducted, jointly, by India and the USSR. They are known by their acronyms (*i*) ISMEX which stands for the Indo-Soviet Monsoon Experiment of 1973 and (*ii*) Monsoon-77.

The 1973 and 1977 experiments revealed that there was indeed a preferred zone where the monsoon air from the southern hemisphere crossed the equator during its northward traverse towards India. This was the region off the coast of Kenya. The cross equatorial flow was particularly strong in this region, but it was only 1.5 to 2.0 km above the surface. The reason why this should be a preferred zone for strong cross equatorial flow is still not well understood. Some evidence was found after the experiments of 1973 and 1977 that fluctuations in the intensity of the low-level cross equatorial winds were reflected in the fluctuations of rainfall over Maharashtra. A feature of the 1977 experiment was the organization, for the first time, of upper air observations over the Bay of Bengal.

Global Atmospheric Research Programme and MONEX

In response to a resolution adopted by the United Nations in the sixties, the World Meteorological Organization (WMO) and the International Council of Scientific Unions (ICSU) organised a Global Atmospheric Research Programme (GARP) in 1969. This was known as GARP. Under the aegis of this programme, a Global Weather Experiment was conducted for one full year beginning on December 1, 1978. It was one of the biggest ever international experiments—on a global scale—for observing the earth's atmosphere. Different facts of the atmosphere were observed and measured from land and ocean based data collection platforms, and by weather satellites, which now monitor the restless atmosphere continuously over any part of the earth. The programme was launched after several years of intensive preparations and planning. Some idea of the dimensions of the experiment may be gleaned from the fact that in May of 1979 as many as 52 research ships were deployed over the tropical oceans between 10°N and 10°S, while 104 aircraft missions were successfully completed over different parts of the Pacific, the Atlantic and the Indian Ocean.

Of considerable interest to India was a special programme of the Global Weather Experiment. This was the Monsoon Experiment (MONEX). Its purpose was to study the influence of monsoon winds on the general circulation of the atmosphere. In view of its economic impact, the Indian scientists were naturally interested in improving their capacity to predict the vagaries of this seasonal phenomenon, which occurs year after year over the land-masses of Asia and parts of Africa.

In view of its seasonal characteristics, the monsoon experiment (MONEX) was designed to have three components :

(*i*) Winter MONEX from December 1, 1978 to March 5, 1979 to cover the eastern Indian Ocean and the Pacific along with the land areas adjoining Malaysia and Indonesia.

(*ii*) Summer MONEX from May 1 to August 31, 1979 which covered the eastern coast of Africa, the Arabian Sea and the Bay of Bengal together with the adjacent landmass. It also covered the Indian Ocean in the belt extending from 10°N to 10°S.

(*iii*) A West African Monsoon Experiment (WAMEX) over western and central parts of Africa from May 1 to August 31, 1979.

International MONEX Management Centres (IMMC) were set up in Kuala Lumpur and in New Delhi to supervise the winter and summer components of the experiment. A large number of scientists from different countries came and worked at these Centres to plan and implement this internationl project.

Data collection platforms—Land based

We usually distinguish two types of meteorological platforms. The first are surface and land based platforms, while the second represent space-based platforms.

Among the surface-based platforms, the major contributions to MONEX were :

—Three civilian research aircraft from the USA

—Five research vessels from the USSR

—Four research ships and one aircraft from India, and

—One research ship together with a constant level balloon programme from France.

The five research vessels from the USSR moved along the equator in the form of a moving polygon, with a spacing of

approximately 400 km between two adjacent ships. They reached the eastern coast of Africa, and proceeded thence to the southern sector of the Arabian Sea by the middle of May. Subsequently, they were able to monitor the onset of the monsoon over Kerala towards the beginning of June in 1979. Later they moved over to the Bay of Bengal to study the formation and structure of monsoon depressions.

Valuable data were collected by the three civilian research aircraft from the USA. Of particular interest to India were data on the radiation balance of the earth-atmosphere system over the monsoon regime. Equipment were provided to measure both the incoming radiation from the sun and the radiation emitted by the earth's surface. The latter, in turn, was determined by interactions between the reflectivity of the soil and the overlying clouds. In addition to radiation measurements, the US aircraft provided unique observations on the thermal structure of the atmosphere by dropwindsondes. Dropwindsondes are small instrumented packages containing meteorological sensors and a tiny parachute. After they are dropped from the aircraft, the sensors transmit their data on meteorological elements to the aircraft as they descend slowly with the parachute. The data were automatically fed into an on-board computer on the aircraft. Forty-six scientific missions were flown by U.S. aircraft for the Arabian Sea phase of the experiment, with a similar number for the Bay of Bengal experiment.

For the first time we had Indian ships that were equipped to measure upper winds over the regions surrounding India. This was achieved with the help of balloon borne instruments known as Omegasondes. The passage of the balloon was tracked by a new navigation system based on the point of intersection of radio beams on very low frequencies. The track of the balloon provided, in turn, a measure of the speed and the direction of the wind. In addition to omegasondes, the Indian ships were equipped to measure the incoming and outgoing radiation, ozone, the conductivity of the atmosphere and its turbidity.

France contributed a ship to monitor the path followed by the monsoon air with constant level balloons. These balloons had an inbuilt device that enabled them to fly at a constant altitude above the earth's surface; consequently, their tracks revealed the traject-

ory of the monsoon air. This was particularly relevant because the summer monsoon crosses the equator off the eastern coast of Kenya, and the deflecting force of the earth's rotation vanishes at the equator.

In addition to novel equipment, the observational programmes of the Indian meteorological stations were specially increased and intensified during the monsoon of 1979. Additional stations were set up to measure upper winds. A special network of eight stations was established for measuring different components of radiation with balloon-borne instruments. The instruments were manufactured and assembled at the workshops of the Meteorological Department of India, but some of the equipment were made by the Department of Space and the National Remote Sensing Agency.

An Indian aircraft (Avro-747) was utilised for recording meteorological observations. This aircraft, which was fitted and equipped for the experiment in very quick time under difficult circumstances, recorded several meteorological elements of which the prominent ones were :

—total air temperature
—pressure
—dew point
—liquid water content of the atmosphere
—radiometric surface temperature and
—radio altitude.

Some of the data were obtained under difficult conditions of turbulent and disturbed weather associated with depressions and tropical cyclones. These observations owe a good deal to the skill and courage of the crew.

In addition to routine observations, a programme for measuring meteorological elements near the earth's surface was designed, jointly, by the Indian Institute of Science, Bangalore and the U.S. MONEX Project. This programme achieved a series of interesting observations at Digha, a coastal station near Contai in West Bengal. The observations were recorded on a 10-metre high mast and they represented, for the first time, meteorological conditions close to the earth's surface under Indian conditions.

There were two other programmes which further enhanced the overall importance of MONEX. The first was designed to investi-

gate the lower stratosphere with the help of rockets launched from (i) Thumba, (ii) Shriharikota and (iii) Balasore with a frequency of approximately one rocket per week. This was organised by the Indian Space Research Organisation (ISRO). The importance of this programme was that it enabled scientists to estimate the response of the upper atmosphere to the lower atmospheric monsoon circulation.

The second programme was concerned with oceanography. Vertical profiles of ocean currents and other oceanographic elements were measured by all ships that took part in this experiment. The data, apart from being of interest to oceanographers were also of interest to meteorologists because they provided information on how the ocean and the atmosphere were coupled to each other.

Space Platforms

Weather satellites are now capable of measuring an impressive array of meteorological elements. Television cameras fitted on these satellites observe the clouds beneath them and transmit the information automatically to a ground station on the earth. This is referred to as an Automatic Picture Transmission (APT) system. The receiving point on the earth is an APT station. A network of eight APT stations was set up in India during MONEX.

Weather satellites are of two broad types : (a) those which orbit round the earth from pole to pole and (b) those which rotate round the earth with the same speed as that of the earth round its own axis. The latter are geostationary satellites because they appear to be stationary with respect to the earth. Geostationary satellites have the advantage of being able to monitor, continuously, any part of the earth's surface round the clock.

A beginning towards reception of data from a geostationary satellite was made during MONEX. A U.S. geostationary satellite—GOES INDIAN OCEAN—was specially moved to a location on the equator at 60°E to cover the MONEX region. Clouds and cloud clusters observed by GOES were beamed towards Bombay by another geostationary satellite, METEOSAT, which was launched earlier by the European Space Agency. Cloud

structures from GOES were first received at a station named Lannion in southern France and then retransmitted to METEOSAT for onward communication to Bombay. Ground reception facilities were set up at Bombay for this purpose. Facilities were also provided at Dhaka in Bangladesh for receiving cloud images from another geostationary satellite launched by Japan.

A wide variety of cloud structures over the Indian monsoon regime is now being brought to light through satellite cloud imageries. Very recently, equipment have been acquired to intercept the cloud images that are transmitted by the polar orbitting satellites launched by the USA and the USSR with great precision. We reproduce in one of the plates the first arrival of monsoon clouds over Kerala towards the end of May, 1984. As the first Indian geostationary satellite has become fully operational, many possibilities have opened up before Indian meteorologists. It is now possible to obtain cloud images transmitted by the satellites with much greater frequency. Currently, polar orbitting satellite pictures are intercepted twice or thrice a day but a geostationary satellite would enable meteorologists to monitor the movement of clouds and associated rain bearing systems once every half an hour. By tracing the movement of clouds, one could infer what would be the winds prevailing over the upper and lower atmosphere. The satellite could also provide information on diverse features, such as the temperature of the surface of the sea, the water vapour content of the atmosphere and the difference between the incoming solar radiation and the outgoing radiation emitted by the earth's surface. These are exciting possibilities for the future of Indian meteorology.

Data Management

To maximise the data output, the transmission, storage and retrieval of data was planned in three stages :

Level—I : Raw primary data, such as the original records of different sensors.

Level—II : Meteorological variables derived from raw data such as, for example, wind speed and direction.

Level—III : Processed data in the form of charts and weather maps.

Level II data have been now compiled and the International MONEX Management Centre in New Delhi has brought out

several volumes of the data. The final Level III data are under preparation, and will become available soon.

What did MONEX reveal ?

We have described in some detail new instrumental techniques that were used during MONEX. Undoubtedly, they provided Indian meteorologists an opportunity to develop these techniques with their own skill and expertise, but the important question that might well be asked is : What did MONEX reveal ? We will try and explain some of the important features that were brought out by MONEX in the succeeding paragraphs. It is important to emphasize that as the data are still being processed, some of the results will need further research.

The genesis of the monsoon

A few years ago there was a controversy about the origin of the monsoon. Many scientists felt that there was very little movement of air from the southern to the northern hemisphere. Observations of MONEX did not support this view. In fact, the trajectories of balloons that were released by France from Diego Garcia and the island of Seychelles clearly showed that they were driven by the monsoon winds towards the Indian coastline. One of the balloons released from Seychelles could be traced right upto Burma (Fig 7.1).

The balloon tracks have improved our understanding of the dynamics of equatorial motion. The acceleration of air is the result of a balance between forces generated by the gradient of pressure, the earth's rotation and the frictional retardation at the earth's surface. Very little was known before MONEX about air trajectories in equatorial regions where the Coriolis force is negligible. By assuming representative forms of the retarding effect of friction, it has now become possible to evaluate the nature of pressure gradients near the equator because the acceleration of air could be determined by monitoring the path of balloons at successive intervals of time.

A considerable volume of data was collected on the strong cross-equatorial jet stream near the earth's surface off the coast of Kenya. It was discovered for example that the intensity of the low level jet showed strong diurnal variations. The maximum

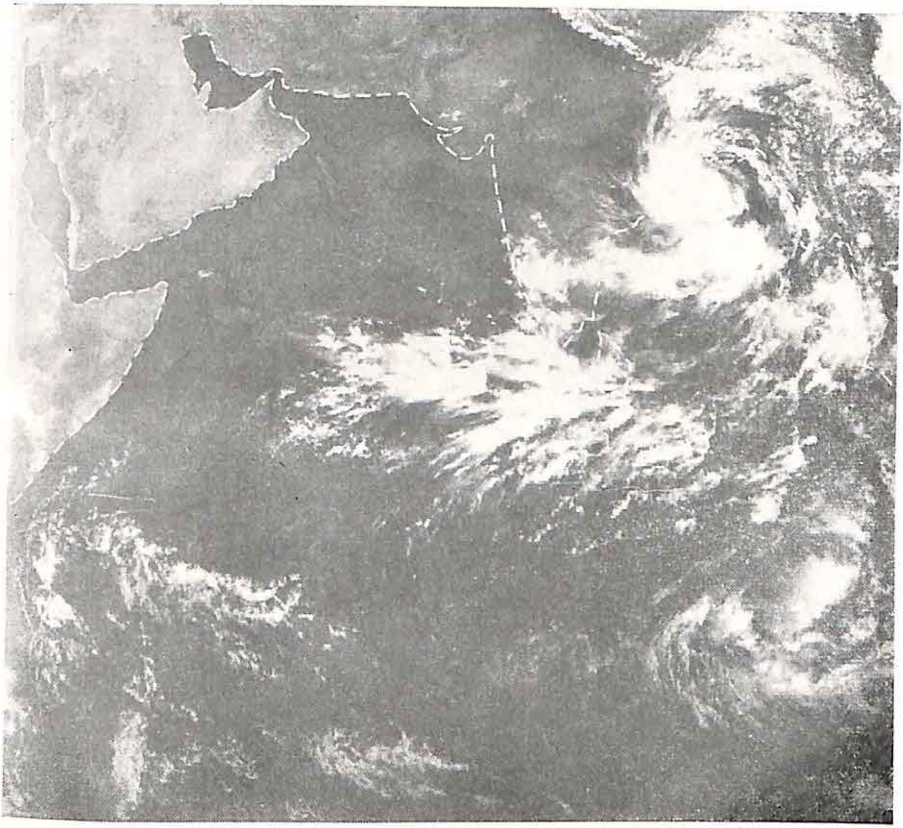

Plate 1 : Monsoon clouds approaching Kerala on June 3, 1982 observed by the Indian Geostationary Satellite. Note the heavy clouding over the north Bay of Bengal, and no clouds over the western part of the Arabian Sea. (Courtesy : Director General of Meteorology, India)

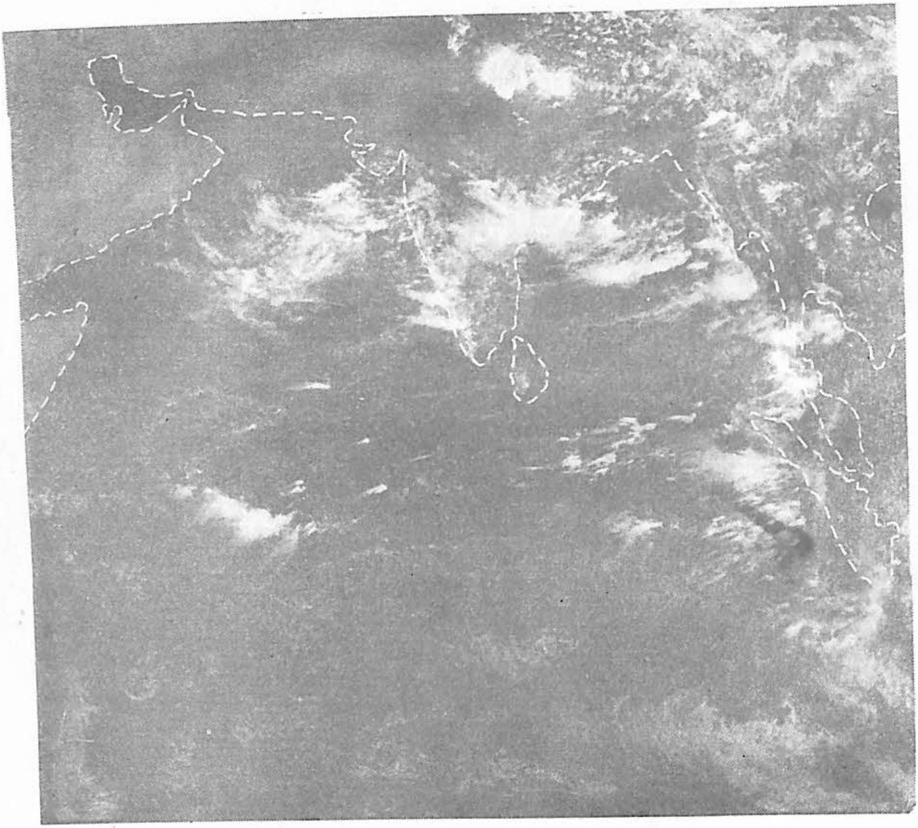

Plate 2 : Monsoon clouds on July 19, 1982. A weak monsoon phase. (Courtesy : Director General of Meteorology, India)

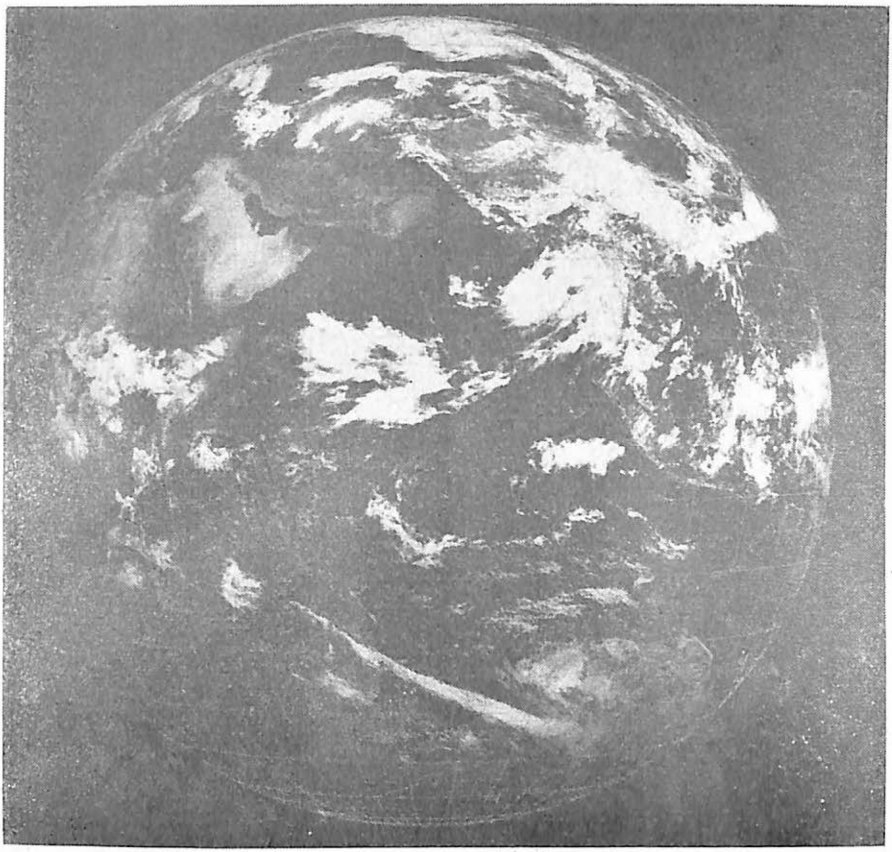

Plate 3 : Onset of the Monsoon over Kerala on May 30, 1984. Note the cloud free zones of northern Arabian Sea and the western sector of the Arabian Sea. The monsoon clouds approach India from the southern hemisphere. (Courtesy : Director General of Meteorology, India)

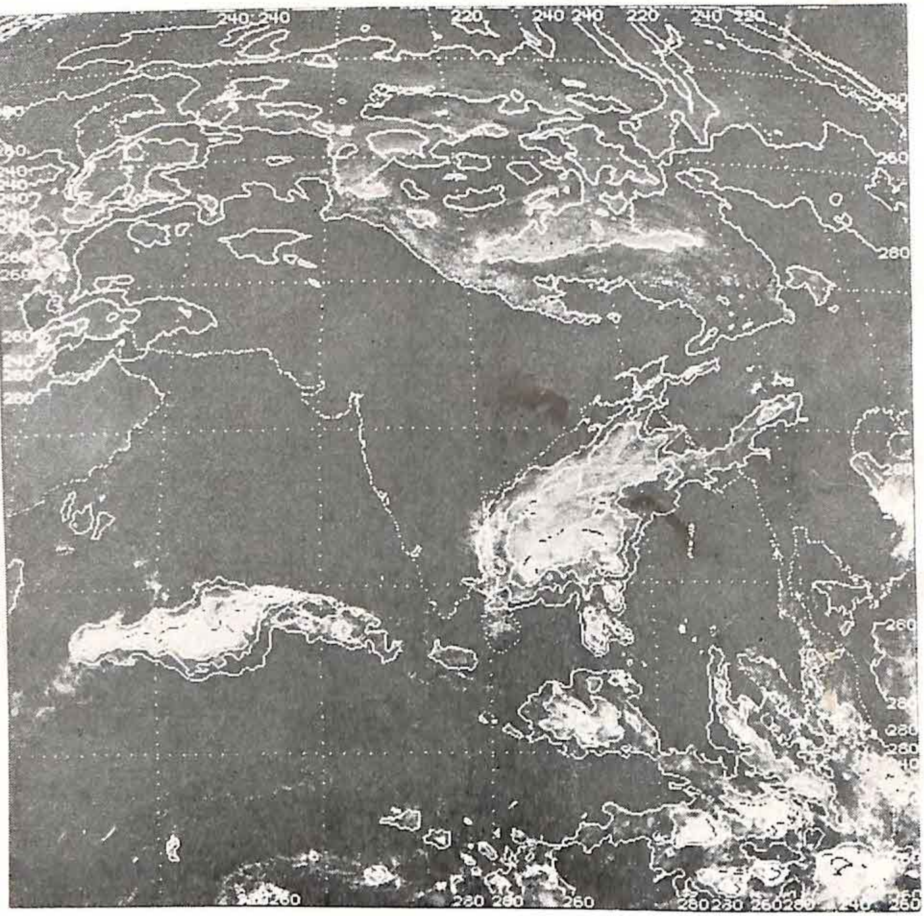

Plate 4 : Lines of equal temperature derived from satellite observations. (Courtesy : Director General of Meteorology, India)

winds during the day were weaker than at night because of convective activity and clouds. Evidence was found of variations in the intensity of the jet, which could be related to surges in the Mozambique channel. Some of the surges could be linked to the passage of low pressure systems.

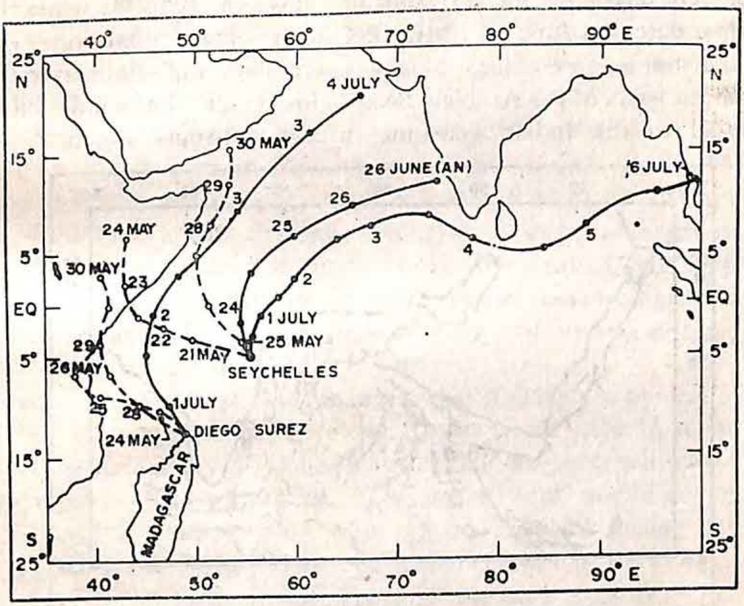

Fig. 7.1—Trajectories of constant level balloons released from Diego-Suarez and Seychelles in the Indian Ocean. Winds before and after monsoon onset are represented by balloon tracks with broken and continuous lines respectively.

This low level jet is strongly coupled to the Somali current and coastal upwelling. It was observed that the region of coastal upwelling tended to move slightly northwards with the progress of the monsoon. Small eddies were discovered that were embedded within the main body of the Somali current. The movement of these eddies emphasized the importance of coastal effects on the formation of the Somali current. Indeed these observations have helped meteorologists to understand the nature of the stress that is exerted by the wind on the surface of the sea. The sea, because of its slower response, is often said to be the

memory of the atmosphere. Consequently, these observations will help us to understand how changes in wind strength determine the extent of coastal upwelling.

Onset of the monsoon

The onset of the monsoon over Kerala in 1979 was delayed by over ten days. It hit Trivandrum only on June 11, while the normal date was June 1. MONEX data suggest that this was caused by a pronounced clockwise circulation of winds over the southern parts of the Arabian Sea. This made the winds blow parallel to the Indian coastline in a north-south direction. If

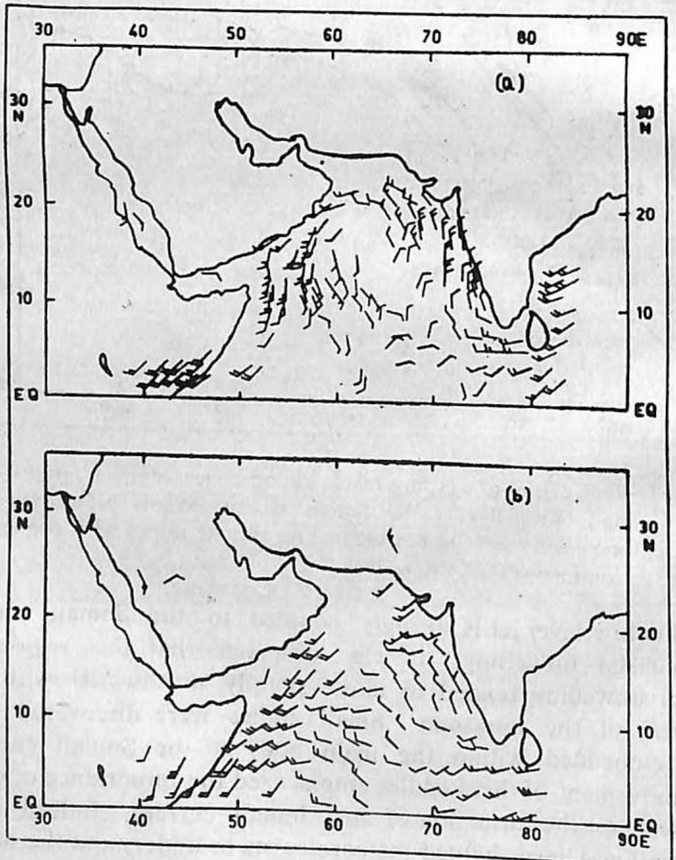

Fig. 7.2—Satellite derived winds over the Arabian sea : (a) before onset of monsoon (June 1, 1979) and (b) after onset of monsoon over Kerala (June 11, 1979).

it was a normal monsoon, the winds should have been hitting the coast in a perpendicular direction blowing from the west to the east. In figs. 7.2 (a, b) we indicate the satellite-derived winds towards the first of June when the monsoon should have set in over Kerala, and around June 11 when the monsoon did eventually set in. The difference between the two wind fields is most pronounced. The first figure reveals clearly the existence of an anticyclonic circulation over the southern parts of the Arabian Sea, while the second figure shows the disintegration of this anticyclone which enabled the monsoon winds to hit Kerala ten days after the normal date. We are thus led to infer that the onset of the monsoon is delayed whenever a large anticyclonic circulation tends to extend northwards over the Arabian Sea. This is an important result, because it will enable meteorologists in future to be on the look out for indications whenever an extension of anticyclonic circulation is likely.

Further evidence of the changes that took place by way of a disintegration of an anticyclonic wind system over the Arabian Sea was provided by the constant level balloons launched by the French contribution to MONEX. This experiment organised by France was performed from May 15 to July 10 during the Arabian Sea phase of MONEX. 88 balloons, which were equipped with temperature, humidity and pressure sensors, were released from the Seychelles Islands and Diego Suarez. The information conveyed by the balloons was obtained by a system of interrogation mounted on polar orbitting satellites (TIROS-N and NOAA-6) launched by the USA. Their positions and the data were received on almost real-time.

Fig. 7.1 depicts the balloon trajectories before the onset of the monsoon of 1979 and after its onset. Prior to the onset of the monsoon the balloons from Seychelles and Deigo Suarez hit the eastern coast of Africa. One of them even flew northwards to strike the Arabian coastline. But subsequent to the onset, most balloons went towards the Indian coastline, and as mentioned earlier, one of them was traced even up to the Burmese coast. The evidence indicates that there is some mechanism by which the onset of the monsoon is triggered by an alignment of wind and pressure systems in the southern hemisphere, which results in a preferred zone of strong cross equat-

orial flow off the east coast of Africa. A feature which has received considerable attention is the formation of a vortex in the south-eastern sector of the Arabian Sea, almost simultaneously with the arrival of the monsoon. Computations reveal an increase in the kinetic energy of the atmosphere with this vortex formation. During the monsoon of 1979 this vortex was first located on June 14, while the arrival of the monsoon took place on June 11. It is not yet clear whether the vortex is necessary, for it is not observed in other years. But, the increase in kinetic energy was observed even in the years after MONEX.

The earth-atmosphere radiation balance

When we consider the radiation balance of the earth-atmosphere system, two important components stand out. One is the downward solar radiation from the sun in the short wave-lengths from 0.2 to 4.0 microns. The second important component is terrestrial radiation emitted by the earth's surface. This radiation is within the limits of 4.2 and 80 microns.

The radiational cooling or warming of a slab of the free atmosphere is measured by the difference between downward solar radiation and upward terrestrial radiation. Unfortunately, it is not simple to evaluate this difference, because both the incoming and outgoing radiation are depleted, and modified by other constituents of the atmosphere. The principal atmospheric constituents which absorb energy from both solar and terrestrial radiation are water vapour, carbon dioxide and ozone. The absorption of ozone is largely confined to the stratosphere, which is the region above 16 km in the atmosphere over India.

There are three other features that are important for the radiation balance. The first is the distribution of clouds. Clouds not only reflect solar radiation but also emit longwave radiation like the earth's surface. Secondly, a good part of the solar radiation is scattered by large aerosols and dust particles. This is important for the semi-arid parts of northwest India and the countries of West Asia, because a heavy dust load exists in these regions. Estimates suggest that for every square mile of north India, there are about 5.5 tonnes of fine dust suspended in the air. Lastly, the reflective properties of the land and the sea have

an important feed-back on the radiation balance. The reflecting power of the earth's surface is known as its albedo. It varies widely over different sectors of the monsoon regime. Generally, desert sands or snow-covered mountains are much more reflective than wet soil or the sea. For many years, the balance between the incoming and outgoing radiation was imperfectly understood. But, during MONEX we had observations of both the incoming and outgoing radiation by instrumented aircraft over different layers of the atmosphere. Fig. 7.3 shows two profiles of temperature changes caused by radiative warming and cooling over Saudi Arabia, recorded by US aircraft. Profile A is a typical afternoon ascent, while B refers to the morning profile. The latter shows a general cooling of the atmosphere from 1 to 9 km (900-300 mb), while the former is indicative of a warming of the atmosphere from 2.5 to 7.5 km (750-400 mb).

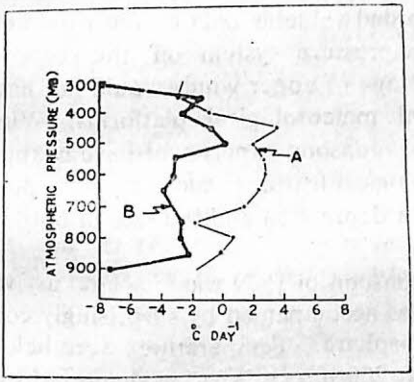

Fig. 7.3—Heating Profiles over Saudi Arabia (22°N 50°E) A—Afternoon and B—Morning.

Radiation measurements over India have revealed another interesting feature. An inverse relationship was observed between the intensity of the outgoing radiation in the lower atmosphere (surface to 6 km) and the middle atmosphere between 6 and 12 km. Clearly, this indicates a difference in the structure between the lower and the upper atmosphere during the monsoon. It suggests that whenever there is a decrease in outgoing radiation in the lower stratosphere, it is accompanied by an increase in the outgoing radiation in the upper atmosphere. This inverse association must be accounted for by the distribution of clouds. In the lower atmosphere the cloudiness of the monsoon circulation prevents outgoing radiation, but in the upper atmosphere it leads to an increase in outgoing radiation because the cloud tops act as radiators, just as the earth's surface.

Rain Bearing Systems

During the monsoon experiment, data were collected on the rain bearing systems of the Indian monsoon. Wind profiles were recorded for a depression in the Bay of Bengal during the first week of July 1979. The data collected by the Indian and the US aircraft revealed, for example, considerable temperature gradients prevailed over the field of a monsoon depression. The existence of such gradients was not known before the Monsoon Experiment. The Indian aircraft also recorded valuable data on the wind field around a mid-tropospheric low pressure system off the coast of Gujarat. The extensive coverage of upper winds would not have been possible with conventional meteorological platforms. Wind and temperature data from the monsoon experiment have enabled us to compute several meteorological features, such as the speed of rotation of the air around a depression and the rate of outflow of wind.

A peculiar feature of the monsoon of 1979 was its late arrival and early withdrawal. This was accompanied by surprisingly cool temperatures in the upper atmosphere. Temperatures were below normal by as much as 4°C at 200 mb (12 km) in the months of May and June of 1979 over north India and most parts of southeast Asia. Low temperatures also prevailed for some periods in July and August, and they were associated with stronger than normal westerly winds. The late arrival and early withdrawal of the monsoon in 1979 led to large rainfall deficiencies over many parts of India. The average rainfall deficiency for the country as a whole was 16%, but in some meteorological divisions it was as large as 51% for the period of June to September, 1979.

Monex led to a new thrust by Indian meteorologists on the design of numerical models to simulate different facets of the monsoon. The advantage of mathematical models is that they enable us to quantify the different physical processes that make up the monsoon. This will be discussed in the Chapter VIII.

CHAPTER VIII

NUMERICAL MODELS OF THE MONSOON

"*Conceivably, Nairobi and Poona meteorologists exchanging visits make better sense than spending time at great computing centres where they may learn to perform minor cosmetic surgery on the latest nine-level primitive equation general circulation model*".

(*From Monsoon Meteorology by C. S. Ramage, 1971*)

It is appropriate to begin this Chapter with the above quotation from Professor Ramage's book on Monsoon Meteorology, because we wish to present a different view.

Meteorologists are concerned with observations. These observations are largely synoptic. "Synoptic" means that the observations refer to a specific time of observation. Consequently, one hears of synoptic observations, synoptic meteorologists and even synopticians. The idea is to represent a collection of information pertaining to a given time. It reflects an average view of weather over an extensive area of the earth. A map depicting weather conditions at a given time would be a synoptic chart to a meteorologist. To assess the changing pattern of weather, several sets of synoptic charts are prepared every day at a modern meteorological office.

On a regular basis, a weather map for the entire northern hemisphere would contain synoptic data from over a thousand observatories spread over the different parts of the world. To obtain a three-dimensional picture of the atmosphere, the data

obtained by vertical probes of the atmosphere are also depicted on maps that pertain to different atmospheric levels. On an average, such maps contain the information transmitted by over 600 balloon-borne instrumented packages every synoptic hour. The task of assimilating such a large volume of data from weather maps is not easy.

Relatively, large numbers of staff are required to operate a meteorological service for a country of the size of India, which has over 500 observatories and another 30 stations making vertical probes of the atmosphere by balloon borne instruments twice every day. The total staff required to just observe the weather at four principal synoptic hours of the data runs into several thousands.

The main question with which we have to contend is how to anticipate the change in weather from this enormous mass of data ? Before the Second World War, it was customary for trained meteorologists to draw on the weather maps lines of equal pressure, of equal temperature, of wind speed and direction and so on to identify the weather generating systems that were likely to affect his region of interest. By this means the meteorologist was able to assess areas of low pressure or regions of disturbed weather generated by tropical cyclones and similar systems. But, meteorologists had little by way of scientific guidelines or principles to understand, for example, why weather systems moved along preferred tracks at certain times of the year, or why they often deviated from these tracks with a rapidity that surprised even the most experienced synoptic meteorologist. Briefly stated, there was little by way of science or quantification behind a weather forecast, apart from the meteorologist's personal experience and subjective judgement. This often created anomalous situations when a sequence of events on a weather map could be interpreted in different ways, depending on the preference of the meteorologist who was assigned the task of analysing the data.

The growth of numerical methods of weather prediction owes its origin to an attempt to assist the struggling science that meteorology was before the Second World War.

In one of his publications, the late Professor Jule Charney, a leading figure in the world of numerical weather prediction,

recalled a meeting of mathematicians at Princeton University towards the end of the War. On that occasion two important mathematicians spoke on the future of computers. Professor Hermann Weyl expressed the feeling that new developments in mathematics were in danger of becoming exhausted unless it received external help, even if it be by "such devilish devices as high speed computing machines". Professor Von Neumann, the other speaker, was enthusiastic enough to feel that the computer had the capability of attacking a wide variety of problems for which no formal solutions were available. He cited the example of weather prediction which was eminently suited for computer technology, because it appeared to him to be a determinate problem.

Unfortunately, Professor Von Neumann's premise concerning the deterministic character of weather change has been a matter of controversy and debate. It would not be without interest to see what the difficulties are.

If we look at a weather chart we see a series of undulating lines resembling waves hurtling through space. That the fluctuations of weather are the result of travelling waves has been the major foundation of weather prediction over many years. But, the meteorologist's difficulty is that it is not always possible to delineate the type of wave we are dealing with because of differences in time and scale, and interactions between the waves themselves.

The atmosphere, as we know it, could support many types of waves. A large explosion, for instance, could set up oscillations which have a period of few minutes. There are tidal waves set up by the attraction of the sun and the moon. Their period is of the order of a few hours. The meteorological waves, which bring about changes in weather, have periods ranging from a single day to several days. One of the central problems of a mathematical model for the atmosphere is to filter out the short period unwanted waves, but to retain the meteorologically significant waves. If we were to increase the scale further and consider changes in climate, one might well think of waves whose periods are of the order of decades or hundreds of years.

Let us confine ourselves to changes in weather on a time scale extending over a few days. If such changes were brought

about by waves moving through a fluid, then the problem should be determinate because the physics of wave motion has been known over many years. The realisation that weather prediction could be based on the principles of fluid motion could be traced back to the well-known Norwegian Meteorologist, Professor V. Bjerknes, around 1904, but the problem as scientists saw it, was indeed so difficult that no serious attempt on this kind of approach was made until the First World War. Around 1920, L. F. Richardson, a British mathematical physicist, devised a prediction scheme for computing the next day's weather. He felt that an amphitheatre with 64,000 men armed with calculators would be needed to calculate tomorrow's weather by solving the mathematical equations that govern the motion of air. Unfortunately, his attempt ended in failure because he had not filtered out the unwanted spurious waves in his model. The disappointing nature of his results discouraged further attempt in this direction till the beginning of the Second World War around 1939. At about this time, there were three important developments which led to a renewal of interest. The first was a demonstration by the celebrated Swedish Meteorologist, C.G. Rossby, that it was not necessary to consider all the imponderables of atmospheric motion. If it was assumed, for example, that the atmosphere was a homogenous fluid which moved horizontally on a sphere, and to which no mass was either added or taken away, then it was possible to design a model atmosphere that would be able to capture a good part of the physics of wave motion. Enlarging on this work, Professor Charney in the USA and Academician Obukhov in USSR were able to derive a consistent set of mathematical equations for weather prediction. Professor Charney was able to show that these equations would eliminate the cause of Professor Richardson's unsuccessful attempt, if constraints were imposed on the equations that govern atmospheric motion.

Lastly, the arrival of the electronic computer was in itself a remarkable event. Professor Charney's scheme was worked out on one of the earliest computers in the USA, namely, the "Electronic Numerical Integrator and Computer (ENIAC)". ENIAC was built as a special purpose computer at the University of Pennsylvania. It was later provided with a general purpose control unit after the design of Von Neumann. To the surprise and

delight of many meteorologists of that generation, the prediction schemes designed by Professor Charney yielded results that showed reasonable agreement with reality.

It is important to realise that a high speed computer is only capable of working out simple repetitive tasks that are prescribed for it to perform. The programme for the computer has to be drawn up by a skilled programmer working in close collaboration with a scientist who knows what he is about to solve. A scientist usually replaces complicated mathematical techniques by simple arithmatical operations like addition, subtraction, multiplication and division. These operations have to be carried out many millions of times. Such operations replace mathematical techniques in an approximate way. A certain amount of in-built error, which is often referred to as a truncation error, is thus a basic shortcoming in numerical weather forecasts.

Usually, the region for which a forecast has to be worked out is covered by a grid, and because of its mathematical simplicity the shape of the grid is usually rectangular. The grid consists of a large number of nodal points. The forecasting equations are then solved by the computer at each nodal point for small periods of time. What the computer does is to use the prediction equation to work out what the weather will be for a short time ahead, say, 10 minutes. It then goes back to the same equations and recalculates what the weather will be for the next 10 minutes and so on until a 24, 48 or 72 hour forecast is obtained. One might well ask : Why calculate weather every 10 minutes ? The answer to this lies in the fact that mathematical equations that govern changes in weather are so intricate that they have no formal solution. One can only unravel their mystery by brute computing power, if very small time intervals are used ; otherwise, we cannot replace the mathematical formalism by the simple arithmetic of addition and subtraction or multiplication and division.

In the middle latitudes, numerical weather prediction has made great strides, its performance is now comparable to forecasts prepared by manual means. On a large number of occasions, machine forecasts prove to be even better than the subjective ones. This has led to considerable saving in manpower and money. But progress in the tropics has been slow. This is because

the dynamics of wave motion is different for low latitudes. The effect of earth's rotation is small and weather systems in the tropics obey different laws. One peculiar feature of the tropics is that it tends to become a region where there is little penetration of weather systems from middle latitudes into the tropics or from the tropics into the mid-latitudes.

Notwithstanding this difficulty, the advent of computers in the developing countries of the tropics has enabled its scientists to design mathematical models. These models use laws that govern the motion of a fluid, such as, air, when it is subjected to external forces. The response of the fluid to different forms of heat exchange, that is, between the land and ocean and the overlying air, or through the release of latent heat, is incorporated in the model by a law which asserts the conservation of energy. What the model does is to assert that any variation in the energy of a parcel of air that may be brought about by a transformation during its traverse over land and sea must be equal to what it receives from its environment. If, for example, through the conversion of water vapour to rain, a certain amount of latent heat is released in the atmosphere, the model will transform this additional energy into the energy of motion. The advantage of a model is that it enables one to perform control experiments on the atmosphere. One could study, for example, what happens to monsoon rains if there were large-scale denudation of forests or if the Arabian Sea was much cooler than usual. Models indicate that tropical weather systems are generated largely by variations on the earth's surface, that is, the weather systems are forced by changes in boundary conditions.

The aim is to mimic the monsoon by a model which is close enough to the real monsoon. Unlike the natural monsoon however, one can deliberately change the forces that control the model monsoon, such as solar radiation or the temperature of the sea surface.

In India, a beginning was made by Das and Bedi to study the impact of topography and solar radiation on the monsoon. A model was designed which starts with an idealised initial state (fig. 8.1). The initial state is a system of winds flowing from the west to the east, but with a reversal in the direction of flow at 5 km. This reversal of wind direction is a common feature of

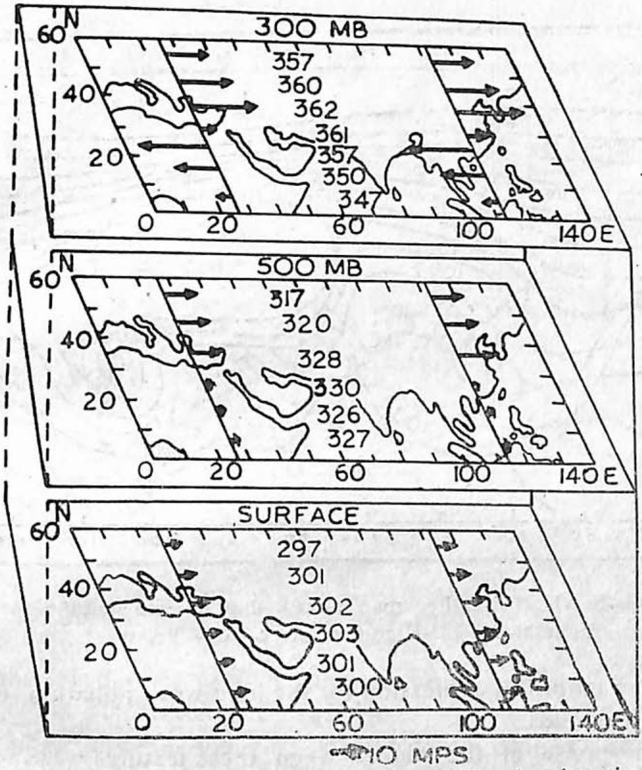

Fig. 8, 1—Initial state of monsoons with meridional and vertical shear. Arrows indicate wind directions and figures represent temperatures (°K).

the Indian monsoon. Temperatures that are typical of the monsoon atmosphere are shown in this figure. We now disturb this initial state by inserting mountain barriers. The response of the model is then computed by a computer. After eight days of simulated time, the model response is shown in fig. 8.2. The interesting feature to note is that the mechanical effect of mountains is not sufficient to generate a monsoon trough.

Next, in addition to mountains we add radiative forcing in the model. This is made up of :

—incoming solar radiation

—effective outgoing radiation emitted by the earth's surface, including the different reflective powers of semi-arid regions and ground with a cover of vegetation ;

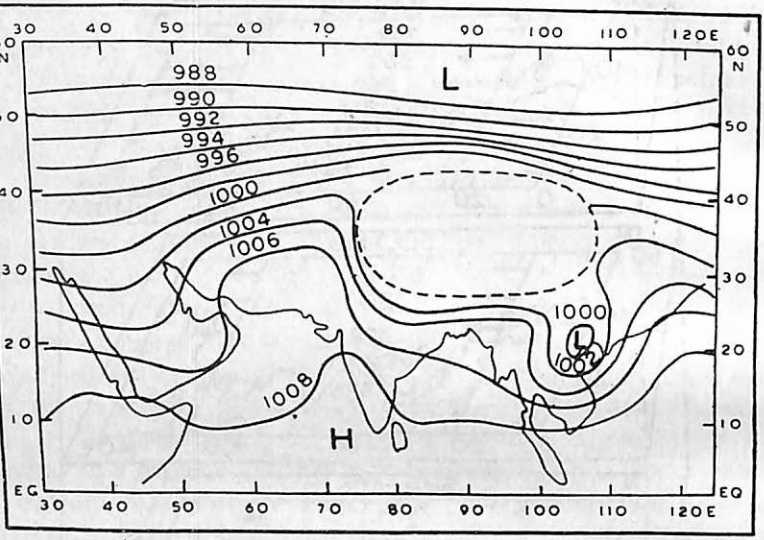

Fig 8.2—Sea-level weather map after 8 days of simulated time with only mountains. H—High Pressure L—Low Pressure

—an empirical expression for the long wave radiation emitted by clouds.

The response of the model when these features were added is shown in fig. 8.3 after ten days of simulated time. The similarity of the model output with the monsoon trough is good.

This is an interesting result because, as we have seen earlier, the location of the monsoon trough is a good indicator of short period rainfall variations. The model experiment suggests that prediction of short-term (5 to 7 days) rainfall variation might become possible if we could monitor the radiation profiles over the country. More refined models are now being designed by scientists who have access to large computers; the refinement is usually in two directions. Firstly, larger computers enable scientists to put in more physics into their models. The Indian model, which we have described, does not have any moisture or rainfall so far, because our intention was to study the feedback from topographic barriers and from vertical radiation profiles. Larger computers also enable one to simulate the three-dimensional structure of the atmosphere in greater detail. What is done is

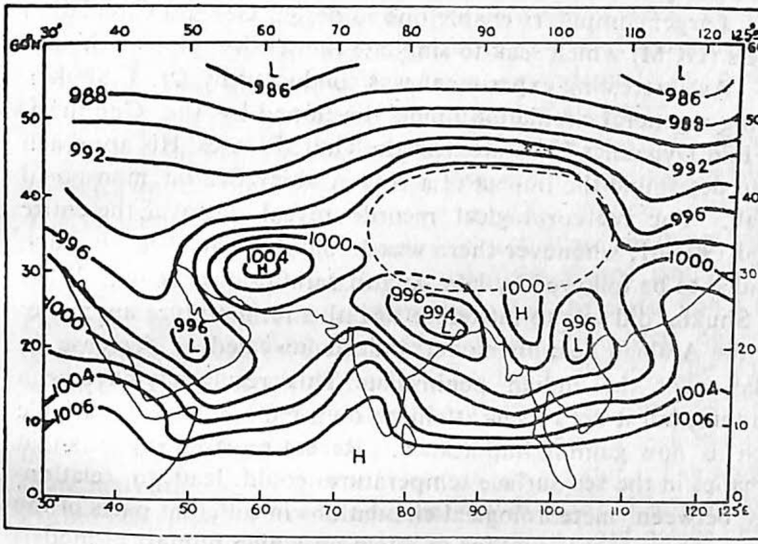

Fig. 8.3—Same as 8.2 but with mountain and radiative effects.

to divide the model atmosphere into a number of layers and assume that the meteorological variables, such as the wind, the temperature and the moisture content do not change within each layer. If we can design a model with a very large number of layers, then it is possible to simulate the vertical structure of the atmosphere with reasonable accuracy. This is important because the temperature, moisture and wind profiles show many changes within atmospheric layers of small depth.

It will be hardly realistic to assume a uniform atmosphere for the model. The atmosphere immediately above the earth's surface, whether it be over land or over the sea, must take into consideration additional features, such as, friction, evaporation from the underlying surface and the flux of heat emitted by the earth's surface. In many countries scientists are able now to design models with greater refinement. A novel feature of these models is parameterization of atmospheric convection and Cumulus clouds. This is clearly of great relevance to tropical modellers.

Although recent models succeed in reproducing broad features of the monsoon, the monsoon trough is not so well captured by

models designed by others. The Indian model appears to perform better. Larger computers enable one to design General Circulation Models (GCM) which seek to simulate the entire global circulation. An interesting experiment was conducted by Dr. J. Shukla, utilising a general circulation model developed by the Geophysical Fluid Dynamics Laboratory in the United States. His approach was to determine the impact of a cold Arabian Sea on monsoonal rainfall. The meteorological records reveal that over the entire period 1901-61, whenever there was a cold Arabian Sea in July, it tended to be followed by late August rainfall over India. What Dr. Shukla did was to introduce negative temperature anomalies over the Arabian Sea; his model indeed revealed a decrease in rainfall over the Indian peninsula. This result may have been fortuitous, but it does focus attention on a meteorological variable which is now gaining importance. Recent research suggests that anomalies in the sea surface temperature could lead to relationships between meteorological circulations in different parts of the globe. There is still another problem on which numerical models are beginning to provide interesting results. For many years, scientists wondered why subsiding motion was observed over the semi-arid and desert lands of the world. It is extremely difficult to measure the rate of ascent or descent of air over large areas. We can hardly expect to measure this by observations of upper winds, because the rate of ascent or descent is such a small quantity that instrumental errors in wind records would suppress the very element which we wish to measure. To get at meaningful results, we would need a very high degree of precision.

To get over this difficulty, meteorologists have tried to use another device. When we consider the movement of air over large distances, it is observed that we can often anticipate their movement by an important conservation principle. This is the conservation of vorticity or spin.

If we proceed on the assumption that air moves only by conserving its spin relative to the earth, then it is possible to infer where the air should rise and where it should subside. Under certain circumstances, it is easier to compute the spin possessed by an air particle, if there is an adequate coverage of upper air soundings over the region under our consideration. The calculation of spin is not so sensitive to the accuracy of aerological

soundings, especially if we consider averages or mean conditions for a particular month or season. The existing network of upper air stations in India and the neighbouring areas of Southeast Asia, the Middle-East and China is just about adequate to make a computation of the type which we are now considering.

There is an important assumption which we have to make if we wish to compute the value of spin from upper air soundings. This concerns an assumption on geostrophic motion. The word "geostrophy" was coined from a Greek word meaning "earth-turning". It implies a balance between the forces generated by gradients of pressure over different parts of the earth, and a force of deflection caused by the earth's rotation. Winds that obey this balance are geostrophic winds. The force due to the earth's rotation is very small in the tropics and vanishes, theoretically, at the equator ; consequently, geostrophic motion becomes increasingly inaccurate as we proceed towards the equator. The accuracy of these calculations would progressively decrease at low latitudes. But, the assumption of geostrophic motion need not be entirely unrealistic if we confine ourselves to consideration of mean conditions. There is no reason to suppose that the motion of air deviates systematically from geostrophic balance, when we consider averages over long periods of time, north of latitude of 10°N for instance.

Let it not be imagined though that with constraints of the type we have just discussed, the computation of spin or vertical motion is an easy proposition. Even for small regions of the globe this involves finding the solution of a set of mathematical equations on a fairly sophisticated electronic computer of large memory.

With the help of this technique, the author studied the circulation patterns for July, a typical monsoon month in India. The results showed an interesting feature. It was found that the monsoon circulation over India was dominated by two principal zones; one of which represented rising air and the other an area of subsiding air. Evidence of strong ascending currents of air was found over northeastern India just south of the Himalayan barrier. On the other hand, there was an equally pronounced zone of descending air over northwest India, especially near the area of the Rajasthan desert. This is shown in fig. 8.4. We

will discuss this in more detail in the next chapter, when we consider the problem of the Rajasthan desert.

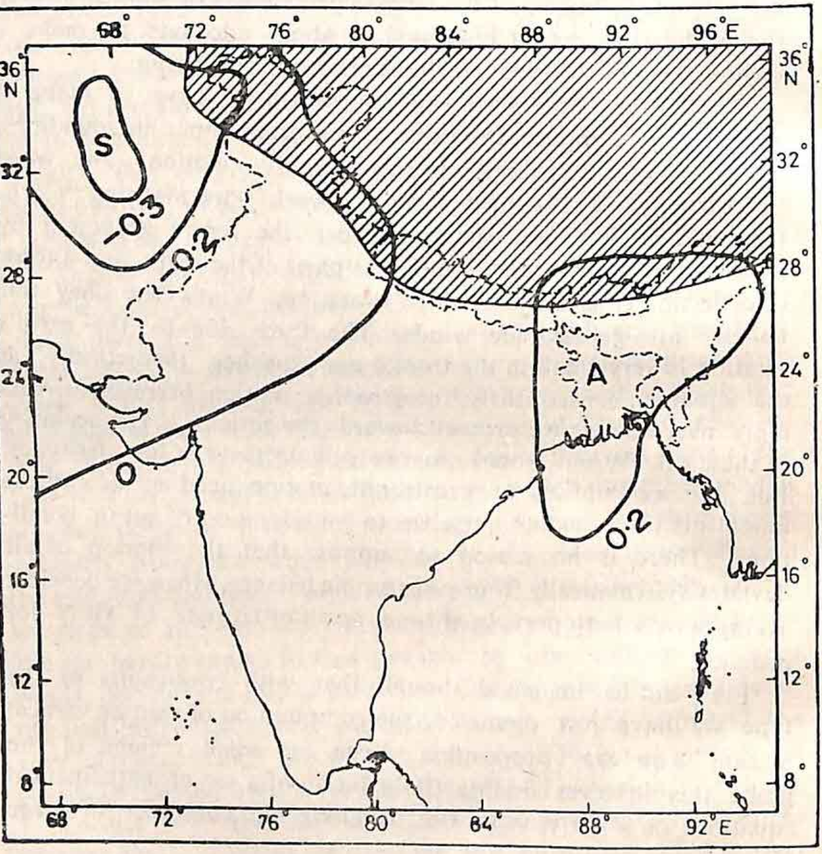

Fig 8.4—Mean vertical motion over India in July at 6.0 km above sea-level Zone of ascent (A) and subsidence (S). (Figures represent vertical motion in cm. sec^{-1}.)
The territorial waters of India extend into the sea to a distance of twelve nautical miles measured from the appropriate base line. Responsibility for the correctness of internal details shown on the map rests with the publisher.

Several experiments with General Circulation Models (GCM) have been devised in the United States. These models portray the three-dimensional structure of the entire atmosphere, as compared to smaller models whose interest lies in the regional aspects

of the monsoon. An early experiment was concerned with the impact of mountains. If all mountains were completely removed from the monsoon regime, the model output suggests that the monsoon circulation would extend much further to the north. In the simulation without mountains, the transition from spring to summer was gradual and the onset of the monsoon was delayed. There is little doubt that this work suggests very strongly that mountains are capable of modulating different aspects of the monsoon circulation. There seems to be an association between mountains on the one hand and thermal characteristics of the monsoon air on the other. This association is often disturbed by different types of clouds, radiation imbalances and sea surface temperatures. Further modelling experiments are needed to improve our understanding of these aspects.

Of considerable interest is yet another experiment. The experiment began with conditions that prevailed over the earth in the ice age about 18,000 years before the present date. This was done with the help of changes in the orbital characteristics of the earth. The experiment suggested that a weaker summer monsoon circulation must have prevailed in the ice age than at present. It leads one to believe that this was primarily caused by the increased reflectivity of ice sheets over south Asia rather than by the colder sea surface temperatures during the ice age. Some support for this conjecture is found by a negative association that is often observed between the extent of snow cover over Asia and the amount of monsoon precipitation in the following summer.

We began this chapter with a quotation from Professor Ramage's book on 'Monsoon Meteorology'. In the first part of his statement Professor Ramage suggests greater interaction between the tropical meteorologists in Nairobi and Pune. There can be no doubt about the need for greater exchange of ideas and views between meteorologists at different centres within the tropics. But, in the second part of his statement he suggests that tropical meteorologists need not concern themselves with minor "cosmetic surgery" on General Circulation Models. In our view, this is a negative statement because, as we have seen, numerical models in the last two decades have very considerably improved our understanding of the physics of monsoons. There is, in our

view, much greater need today for tropical meteorologists to design models, if necessary, at a central location in the tropics, equipped with a good computer and peripherals. This will enable them to experiment and innovate with new physical concepts. What is needed in the tropics today are not men who are merely good in describing a sequence of events, but men who are able to think and apply reasoning to explain why the monsoons behave as they appear to do on a weather map. To meet this objective, there need be no dichotomy between numerical modellers and synopticians, because they both need each other.

CHAPTER IX

THE RAJASTHAN DESERT PROBLEM

AROUND THE END OF the First World War 1918, Dr. Vladimir Koppen of the University of Graz in Austria studied the different ecological formations of the world. He found that nearly 14% of the earth's land receives less than 25 cm of annual rainfall. If we also consider regions which receive only between 25 and 50 cm of annual rainfall, another 14% of the earth's surface may be classified as arid or semi-arid.

These areas, which are often referred to as "Koppen Deserts", may be further sub-divided into twelve major deserts of the earth.

Sahara is by far the biggest desert. It stretches across the width of North Africa. A peculiar feature of this desert is that it contains mountains that are as high as 3.5 km which receive snow, but most of the deserts at lower altitudes receive less than 2.5 cm of rainfall a year. There is evidence to suggest that the hamlet of Dakhla in the Sahara went without rain for a spell of eleven consecutive years.

Coupled with little rainfall, the characteristic features of deserts are high temperatures and low humidity. The highest temperature ever recorded on earth (58°C) was measured at Azizia in the Libiyan part of the Sahara Desert. Temperatures of the order of 49°C are fairly common.

An extension of the Iranian and the Arabian Desert is represented by the Thar Desert of western India and Pakistan. Some people refer to it as the Great Indian Desert, which lies to the east of the Indus river. The annual rainfall over the Thar Desert is less than 15 cm. The area of deficient rainfall surrounding this region consists of most of the western districts of

Rajasthan. The total area of this desert is about a fifth of the total area of India.

For many years it was realised that if we could reclaim this desert, and make it more suitable for agriculture, there would be economic benefits for all concerned. An attractive feature of a project of this nature is that it invites a concerted effort from many disciplines. The task of improving irrigation facilities or the structure of the desert soil are clearly formidable problems in agricultural engineering. For the meteorologist, the area provides interesting possibilities for climate modification.

In the months of July to September the Thar Desert is under the spell of the Arabian Sea branch of the monsoon. The moisture content of the air over Rajasthan is high, but the rainfall figures are disappointingly low. Observations of the temperature gradient over Rajasthan suggest the existence of an inversion in the lower troposphere at about 1.5 km above sea level. This is indicative of an extremely stable density gradient in the atmosphere. Under these conditions the atmosphere tends to suppress large scale ascent of air; consequently, the air is never able to rise to a point of saturation despite its high moisture content. In recent years, meteorologists have tried to study the nature of this thermal inversion. One is interested to find out if conditions could be created which would alter the thermal structure of the atmosphere. If the inversion could be removed, one could conceivably increase the tendency for rising air currents to occur over Rajasthan. As the supply of moisture is already there, this might lead to an increase in rainfall. A chain of events could thus be set in motion whereby an increase in precipitation would lead to greater vegetation and a gradual improvement in soil structure.

The Archaeological History of the Thar Desert

Many historians and archaeologists have drawn our attention to the fact that the Rajasthan Desert was once the centre of a flourishing civilisation. Palaeontological studies by G. Singh in India and by Reid Bryson at the University of Wisconsin, U.S.A. suggest that the history of the area may be divided into four periods :

(*i*) 8000 B.C. : Moist, wet and cool climate.

(*ii*) 8000-3000 B.C. : Dry climate, but less arid than at present. Evidence suggests the beginning of agriculture.
(*iii*) 3000-1700 B.C. : Period of higher rainfall.
(*iv*) 1700-1500 B.C. : Dry conditions. Evidence of fresh water lakes drying up.

Following the last period, the indications are that very arid conditions prevailed over Rajasthan.

From about 2000 to 1700 B.C. this region was the centre of the Harappan civilisation. The relics of this civilisation suggest that the area was favourable for human habitation during this period. An excavation at a site near Kalibangan, near the river Ghaggar, revealed a number of dessicated valleys suggesting a system of rivers that once reached the sea as a tributary of the Indus.

In the next phase, another civilisation known as the Painted Grey Ware Culture flourished in this area. The beginning of this period was around 500 B.C. There seems to have been a break of about a thousand years between the termination of the Harappan civilisation and the beginning of the Painted Grey Ware Culture.

In the last phase, archaeologists refer to the Rangmahal Culture, which flourished in the early centuries A. D. We do not know if there was any break in civilisation between the second and the third phase, but we do find that the history of this area was one of extreme fluctuations. There were periods of intense cultivation followed by years of dry or semi-arid conditions. The important point to note was that the desert managed to survive an unfavourable climate on at least two occasions; on each occasion its survival was succeeded by the birth of a new culture or civilisation.

These fluctuations in climate may be either part of a world wide change in climate, or they could be of local origin, perhaps induced by human activity.

It has been pointed out that plants that grew in this region during the Harappan civilisation are much the same as those found in the fringes of the area today. One is thus led to suspect that these changes were not of world-wide nature.

Example of areas which are now deserts, but which at some

time in the past enjoyed a favourable climate are not difficult to find. The climate of the Sahara was reputed to be wet and moist between 60,000 to 6000 B.C. Geological evidence suggests that sometime after the glaciers of the last ice age began to recede, the climate of the Sahara became increasingly hostile. Around a Sandstone plateau in southern Algeria, paintings and engravings have been discovered which provide glimpses of the lives of tribes that occupied that plateau from 8000 B.C. to nearly the time of Christ. Pictures of long-horned cattle and herdsmen suggest that these parts of the Sahara were once fertile grazing pastures.

It is generally agreed that, if the land that is currently occupied by deserts could be reclaimed, it would be a great step forward. Opinions differ, but the general view seems to be that the Rajasthan Desert is advancing inland at the rate of a quarter to half a mile every year. If we project this rate of advance backwards in time, the origin of the Thar Desert would be between 400 to 1000 years from now.

It is difficult to attribute the progressive decline of the soil to world-wide climatic changes. If large climatic changes were involved, there should have been flourishing periods between the end of the Rangmahal period and the present day. But, there is little or no evidence to support the existence of a flourishing period in Rajasthan during the last 1000 years.

How are deserts formed ?

The atmosphere responds to the external heat supplied by the sun. The intensity of the solar beam at the top of the atmosphere is around 1360 watts for every square metre of the earth. We refer to this as the Solar Constant. It has been conjectured that even a small change in the Solar Constant could lead to changes in climate. But it is equally important to realise that the atmosphere interacts with different features of the earth's surface, such as, the soil, the rocks, ice and water and the climate also depends on the gaseous composition of the atmosphere. Changes in the structure of the earth's surface or the composition of the atmosphere could bring about changes in climate. In view of these complex interactions, it is difficult to find out how the planetary

atmosphere will respond to a large change in external forces. For example, if there was a large change in the solar constant, or if there was a sudden increase in the dust content of the air, how would the atmosphere respond? The models, which we described in the last chapter, suggest that snow cover or the properties of the ocean's surface layer and the internal behaviour of both the atmosphere and the oceans could bring about climatic variations on a time scale of decades. As an example, there is much debate today on the possible impact of an increase in the carbon dioxide content of the atmosphere.

There is a growing body of opinion which suggests that deserts are the results of human misuse of land at times of climatic stress. This could take the form of intense grazing for animal survival in areas that lie adjacent to a desert.

In addition to increase in grazing, there is a tendency to deplete the resources of wood in areas lying on the fringes of deserts. There is evidence to indicate that many parts of north-west Himalayas are constantly eroded by frequent assaults on once prosperous forests.

An interesting meteorological observation is concerned with the preponderance of anticyclones or high pressure cells lying over the principal deserts of the world. The atmosphere below an anticyclone generally subsides and, as a consequence, it is warmed by compression. This is a paradoxical observation because deserts are regions of the earth that are much warmer than the surrounding lands. One would normally expect the air over a heated surface to rise, but actually it is here that the air tends to sink. Apart from anticyclonic circulations, subsiding air may be caused by mountain barriers. To the west of the Andes Mountains there is a region of continuous subsidence over the coastal areas of north Chile and Peru. This leads to the Atacama Desert.

A possible reason for deserts and semi-arid conditions is the marked absence of rain generating weather systems and the absence of humid air streams. But, as we have seen, the absence of humid air is not really a reason for the Thar Desert, because this region is covered by the monsoonal winds every year between June and September.

Models to explain desert formation

In the previous chapter, we discussed how meteorologists were designing mathematical models to investigate different facets of the atmosphere. This approach is interesting for studying specific feedback processes, such as, variations in the reflectivity of the soil or changes in the storage of soil moisture.

Interest in models to simulate deserts was initiated by Professor Jule Charney in the United States. Professor Charney noted that many areas of central and northern Sahara, the eastern part of Saudi Arabia and southern Iraq had a negative balance of radiation at the top of the atmosphere, in spite of an intense input of solar radiation through the cloudless desert atmosphere. Measurements by weather satellites revealed that over desert areas there was more radiation leaving the top of the atmosphere than what was coming in by way of a direct input from the sun. In a sense then, the desert was being cooled because the outgoing radiation was enhanced by very high surface temperatures. The absence of clouds, low humidities and the high reflectivity of the desert sands add to a negative radiation balance. The reflectivity of the soil is known to meteorologists as its albedo. The important point which Professor Charney stressed was : As the desert represents a deficit in the radiation balance of the atmosphere, the air over the desert should get cooled. But this does not happen because the cooled air subsides and subsidence leads to warming by compression. In fact, subsidence makes the desert increase its own dryness. There is thus a biogeophysical feedback mechanism which makes a desert maintain itself. A General Circulation Model (GCM) designed by him showed that a change of albedo from 14 to 35% could lead to sharp reduction of clouds and rains. The net radiation over India measured by weather satellites also indicate a negative balance over Rajasthan and the adjoining parts of the Middle East.

In the preceding chapter we have seen how the monsoon circulation is marked by a region of ascent over northeast India and a zone of subsidence over northwest India. There is thus a pattern of subsidence over the Thar Desert which is obtained by direct computations with the help of prevailing winds and the thermal structure of the atmosphere over Rajasthan in July. An important feature of these computations is that it enables

us to infer at what rate the atmosphere must be warmed or cooled to maintain a steady monsoon circulation over India. On the basis of these computations it was observed that the atmosphere must be warmed at the rate of 3.2°C per day over northeast India and cooled by 2.4°C per day over northwest India, if we are to maintain a steady monsoon circulation. Interestingly, this agrees with our concept of a Walker Cell for the Asian summer monsoon.

A few simple calculations indicate that the warming required over northeast India was entirely reasonable if we consider the heavy rainfall that occurs over sub-Himalayan West-Bengal and Assam. The latent heat released by heavy rain would be sufficient to warm the atmosphere by 4 to 5°C per day in northeast India. The predicted warming was thus of the correct order of magnitude and sign over northeast India.

The predicted cooling over northwest India was not explained so easily. At first sight one might be tempted to reason that this cooling was the result of outgoing long wave radiation. But estimates of this quantity reveal that maximum cooling by this process could only be about 1.6 to 1.8°C per day. How are we to account for an additional cooling at the rate of about 0.8°C per day as predicted by theory ?

Field experiments over Rajasthan revealed an interesting result. It is possible to measure the rate of incoming and outgoing radiant energy at different levels of the atmosphere by a ballon-borne instrument known as a radiometer. Measurements with this instrument revealed that the observed rate of cooling over the atmosphere was very nearly the same as predicted by theory, namely, 2.4°C per day.

The suggestion was made that the additional amount of cooling required by theory could be attributed to the presence of dust over the desert. If this hypothesis was correct, then the presence of dust must enhance the subsidence rate of air over the desert by over 50%.

We have here an alternative hypothesis to that of Professor Charney. Professor Charney's view emphasizes the feedback from a high albedo leading to a radiation deficit over the Thar Desert, but ignores the role of dust. On the other hand, the high dust-load of the air over the Thar could also provide a feedback that leads to a larger negative radiation balance.

The decline in soil texture has led to an abnormal increase in the dust content of the atmosphere over Rajasthan. During the summer of 1966, a number of reconnaissance flights were conducted to determine the extent of the atmospheric dust over Rajasthan. The data collected during these flights were combined with similar observations over different parts of Southeast Asia during the summer months of 1962 and 1963. The overall picture that emerged showed that the dust cover was much more extensive than what had been imagined before.

A similar deep and dense layer has been found to extend from the shores of Sahara in North Africa to Arabia. Thereafter, there was a slight decrease in the density of dust as one approached the southern coast of Iran and the Mekeran Coast; but the density and the depth of the dust increased sharply as the aircraft flew over the arid regions of Rajasthan. Subsequently, the density of dust decreased as the aircraft flew to the south and to the east of the Rajasthan desert. Although not as dense as over the desert, considerable quantities of dust were observed over the industrial areas of northeast India and over parts of Burma, Thailand and Cambodia.

Visual observations of the top of the dust layer suggest that it was around 7 km over the northeastern parts of the Arabian Sea. Thereafter, it increased to about 9 km over the Rajasthan desert, but decreased to about 5 km as one approached the Gangetic valley.

These particles of dust are extremely small. Their diameters are of the order of a micron which is a millionth part of a metre. They are small enough to remain in suspension in the atmosphere for several days. Often the clouds of dust are sufficiently dense to completely obscure the sun even at mid-day. Most of the samples over Rajasthan were found to be particles of quartz, although small quantities of silicates, such as mica and felspar were also observed.

The mechanism which lifts such large volumes of dust to great heights is a matter on which our understanding is not very clear. The Thar Desert is an area where there is large scale subsidence of air. But if this is so, where is the origin of the dust? If it was of local origin, then we need fairly strong currents of ascending air to lift dust particles to a height of 5 km or more.

It seems reasonable to assume that a large portion of the dust over the Thar Desert was transported from the deserts of Arabia to the west of the Thar.

The role of dust in the radiation balance of the atmosphere

We have suggested that the presence of dust increases the rate of subsidence of the desert air by about 50%. If we carry this reasoning a little forward, it implies that in the absence of dust there would be less radiational cooling. As a consequence, there would be less of subsiding air and a deeper monsoon current. But, a decrease in subsidence would mean more ascending air and rainfall because the moisture is already present. The only reason why the monsoon water vapour cannot be converted into rain is the lack of a lifting agent to cool water vapour to its condensation level. If the source of the dust is the desert itself, then the desert would appear to be maintained by a self-sustaining mechanism. The presence of a protective vegetative cover could prevent the dust being raised by the wind and this would arrest the development of a desert climate, which is inimical to the growth of vegetation.

It is natural to enquire why does dust enable the atmosphere to cool at a faster rate. The answer to this question must be viewed in two ways.

The radiation balance of the atmosphere is determined by two dominant processes, namely, the absorption of incoming solar radiation and the emission of longwave radiation from the ground. Hence, we need to examine the role of dust cover both on the radiation from the sun as well as the outgoing radiation emitted by the earth.

The primary effect of dust is to cut out a good portion of short wave solar radiation by scattering. This prevents the solar radiation from reaching the ground. Scattering is a term used by physicists to indicate irregular reflection of beams of radiation striking a particle of dust.

But the influence of aerosols on terrestrial radiation is more difficult to assess. The main difficulty is that dust both absorbs and scatters longwave radiation. There are certain regions in the spectrum of terrestrial radiation where the primary influence is that of gaseous constituents of the atmosphere, such as water

vapour and carbon dioxide. But, there are other regions where the influence of dust predominates. We still do not know for certain whether the primary role of dust is to absorb or scatter longwave radiation.

One could picture a thick layer of dust as having the same kind of effect as a layer of low clouds. The net flux at the top of the cloud is the difference between the upward flux from the cloud top and the downward radiation from the atmosphere above the cloud. As the air above is usually colder, less radiation comes down to the cloud than it emits to space. Consequently, the cloud top tends to lose energy and is gradually cooled.

At its base the cloud receives radiation from the warmer earth and the air below. The base of the cloud, therefore, tends to get warmer. The net effect is then an overall cooling, if the base of the cloud is sufficiently low. Much the same type of situation prevails over a thick layer of dust near the ground. The warming at the base is marginal but the cooling at the top is quite large. The net effect is again an overall cooling. For simplicity, we assume a single layer of dust particles. If there were several layers of dust in the atmosphere, the net effect was not likely to be quite so simple. It is for these reasons that we need to know much more about the distribution of dust over the Thar.

Dust laden air streams emerging from the world's dry lands have been known from very early days. Dust from the Sahara, for example, is carried away by air streams out of North Africa towards the Atlantic. The transport of dust should logically be included in modelling experiments of the future, because as we can see, it could have an important impact on the radiation balance over the Thar Desert.

Economic benefits of dust removal

The importance of agriculture to India's economy is well-known. Agricultural activities provide employment to nearly 70% of India's population. They contribute a very substantial amount to the national income. Even so, the congestion over land is increasing at an alarming rate. The reclamation of desert land could not only add substantially to our total output of food, it could provide additional space for an increasing population.

We have discussed in this chapter the meteorological aspects of the problem of desert reclamation. The problem of stabilising the soil is not likely to be an easy one. To provide a grass cover over land which is now covered with sand would be a problem in agricultural engineering, but if it could be done, the meteorological investigations suggest favourable consequences.

Firstly, with an adequate grass cover there would be lesser amount of dust floating in the atmosphere. This, as we have shown earlier, would reduce the cooling rate required to maintain steady monsoon conditions. It would also reduce the region of subsiding air. A reduction in the average subsidence over the desert could create more favourable conditions for upward motion. This in turn would increase the chances of more precipitation. There are several observations which indicate that the run-off from an intensely grazed river basin is nearly 20% greater than one from which no grazing is permitted. More run-off leads to greater erosion, and this reduces the depth of soil and the amount of water it can hold. The loss of water-holding capacity of the soil has an adverse effect on the growth of vegetation.

There is another prospect with interesting possibilities. If there were less dust particles in the air more solar energy would reach the ground during the day. At night the ground would cool more than before because of the larger solar radiation received during day. The overall effect would be to increase the range of temperature variation between day and night. If we could lower the night minimum temperature and increase the heat received during the day, we create conditions which are favourable for the formation of dew. Dew is an important source of moisture for semi-arid and arid lands.

There are numerous possibilities of this nature which could be made to yield material benefits if handled in a scientific manner. The water vapour that is brought in by the monsoon currents over Rajasthan is a valuable source of moisture. It is but natural that we should seek to utilise this natural resource to our advantage.

CHAPTER X

THE ENERGETICS OF THE MONSOON

THERE ARE TWO aspects of the Indian monsoon which are important for understanding its origin. The first is concerned with the heat balance of the atmosphere, especially over India during the pre-monsoon and monsoon months. We need to know how a region of high temperature builds up over northwest India because this is intimately linked with an area of low air pressure. Secondly, it is necessary to acquire a more quantitative measure of the transport of energy and water vapour in the air associated with monsoons. A systematic attempt to study these aspects has been possible only recently because of improvements in our knowledge of the upper atmosphere. This chapter will be concerned with transport mechanisms, because it is important to know how the monsoon fits in with the global pattern of large scale transfer across different latitudes.

The Earth's Heat Balance

The energy for all atmospheric motions is derived from the sun. The intensity of solar radiation falling on a surface of unit area on the outer fringes of the earth's atmosphere has a mean value of 1360 watts for every square metre. This is known as the solar constant. If the earth had no atmosphere and if it completely absorbed the solar radiation falling on it, the earth's temperature would have a uniform value of 245°A (-28°C). But, solar radiation while passing through the earth's atmosphere experiences many transformations. Consequently, the temperature at the earth's surface is very different from a uniform value of 245°A. To understand these transformations it is necessary to know a little more about the nature of solar radiation.

The spectral composition of solar energy is conveniently expressed in terms of the wavelengths of the constituent radiations, Physicists generally measure wavelengths of radiation in any one of the following units :

1 micron	=	10^{-6} m
1 millimicron	=	10^{-9} m
1 Angstrom (A)	=	10^{-10} m

The wavelengths of radiation encountered in nature vary over wide ranges. The region corresponding to wavelengths between 0.40 to 0.75 microns is the visible range of the solar spectrum. All radiation falling within this region is visible to the human eye. The wavelengths which correspond to different colours, for example, are shown in the following table.

Table 10.1
Wavelengths of Different Colours

Colour	*Wavelength (microns)*	*Colour*	*Wavelength (microns)*
Violet	·430	Greenish yellow	·560
Dark blue	·470	Yellow	·580
Light blue	·495	Orange	·600
Green	·530	Red	·640

The zone of spectral wavelengths less than 0.4 microns is the ultraviolet region. Similarly, the radiation of long waves of wavelengths greater than 0.75 microns is known as the infrared region of the atmosphere.

Approximately 99 per cent of the sun's radiation is contained in the wavelength range of 0.15 to 4.0 microns. Meteorologists refer to this range of wavelength as that of solar radiation. Of the total solar radiation, about 9 per cent occurs in the ultraviolet, 45 per cent in the visible and 46 per cent in the infra-red region.

In addition to incoming radiation from the sun, the earth and the atmosphere also emit radiation. The major portion of

this radiation lies in a range between 4 and 80 microns. We refer to radiation within this wavelength as terrestrial radiation. The essential difference between solar and terrestrial radiation is in their wavelength. While solar radiation is confined to wavelengths shorter than 4 microns, terrestrial radiation is entirely in the infra-red region between 4 and 80 microns.

There are three principal processes which determine the radiation balance of the atmosphere : (*i*) reflection of radiation energy by the earth's surface or by clouds, (*ii*) the absorption of radiation by different constituents of the atmosphere and (*iii*) attenuation of radiation by scattering. It is important to note that the reflection of radiation does not mean depletion of energy. On the other hand, the process of absorption implies an extraction of energy from the supply of radiation emitted by the source. The principal constituents of the atmosphere which absorb radiant energy are water vapour, carbon dioxide and ozone. Each constituent absorbs radiant energy in different intervals of wavelength. Much of our difficulty in understanding the earth's radiation balance could be attributed to the irregular manner in which atmospheric constituents absorb radiant energy.

Apart from reflection and absorption, the third principal process which controls radiation is by scattering small particles or aerosols. The process of scattering is a series of irregular reflections by small obstacles. The law of scattering for very small particles, whose dimensions are small compared to the wavelength of the radiation, is fairly well-known. The shorter the waves, the more easily are they scattered or irregularly reflected by small obstacles; in fact, the amount of energy scattered is inversely proportional to the fourth power of wavelength. This provides us with an explanation of the colour of the setting sun towards the evening. The light from the sun is deprived of a large proportion of its shorter blue waves ; the light which gets through is consequently more red. For this reason, the sky at the time of the setting sun appears to us as red. The effect of scattering has received considerable attention in recent years in view of its important effect on the earth's radiation balance. But, the laws of scattering are more complicated when we consider large obstacles, such as, particles of dust. Some measurements of the dust content of the air have been made over the

arid regions of Rajasthan in India, which were mentioned in the preceding chapter.

The earth's atmosphere is largely transparent to solar radiation, if an allowance is made for small absorption by oxygen and ozone in the stratosphere. But, the incoming radiation from the sun suffers considerable depletion by scattering and reflection.

Let us represent the radius of the earth by a, and the undepleted solar radiation interrupted by the earth by $S \times \pi a^2$ where S denotes the solar constant. If we consider this amount of radiation to be distributed uniformly over the entire earth, then a surface of unit area of the earth receives the following amount of radiant energy (E).

$$E = (S \times \pi a^2) \div 4\pi a^2 = \frac{S}{4}$$

As we have seen earlier, the mean value of S is 1360 watts for every square meter, whence S/4 is 340 watts/m^2.

Of this amount, about 35 per cent is reflected back by the earth's atmosphere, mainly by clouds. Another 15 per cent is lost by scattering and absorption. On an average, therefore, the mean inflow of solar radiation is only about 170 watts/m^2.

Now it may be argued that the atmosphere is neither getting progressively warmer nor cooler over the years. Its average temperature remains much the same. A state of long term balance is thus maintained. And, to maintain this balance the power which must be radiated back to space by the earth and its atmosphere should equal the mean inflow of solar energy.

The transfer problem for outgoing terrestrial radiation is not so simple as that of incoming solar radiation. This is largely confined to the region between 4 and 80 microns, and in this region there are small intervals of wavelength where water vapour and carbon dioxide are strong absorbers.

Another complicating feature is created by the presence of clouds. Low level water clouds are excellent absorbers of radiation. We might well regard an overcast sky as a blanket over the earth's surface. Unfortunately, our knowledge of the distribution of clouds over different parts of the earth is still meagre, but this situation will probably be rectified in the next few years because large volumes of cloud data are now coming in through weather satellites.

Despite these limitations, the problem was approached by Sir George Simpson, over fifty years ago, by dividing the entire spectrum of terrestrial radiation into three regions. He considered (*i*) a region in which water vapour does not absorb any radiation (8.5—11 microns), (*ii*) a region in which water vapour will absorb all radiation (5.5—7 microns) and wavelengths greater than 14 microns and (*iii*) wavelengths in which absorption by water vapour is half-way between (*i*) and (*ii*). Included in (*iii*) was a region around 15 microns, where carbon dioxide is a strong absorber. With a hypothetical model of this kind, Sir George Simpson found that the mean outgoing radiation from the atmosphere was 188 watts/m^2, if we assume that about half the earth is covered with clouds. Despite the assumptions that were made, this figure was within a few per cent of the amount needed to maintain a long term balance.

Simpson's early work was followed by H.G. Houghton with more precise estimates. He observed that from the equator to about 35°N, the outgoing radiation was less than the incoming radiation. North of 35°N, the outgoing radiation exceeds incoming solar radiation. There was thus a net transfer of heat from low to high latitudes through the medium of the general circulation of the atmosphere and the ocean. The problem before us is to find out how much of this transfer is accomplished by the monsoon circulation.

Unfortunately, there have been few systematic attempts to study the heat budget of the monsoon. During the International Indian Ocean Expedition (IIOE) a few results were obtained with the help of research aircraft. The observations for the western and eastern sectors of the Arabian Sea are summarised in Tables 10.2 and 10.3.

Columns 2 and 3 of these tables provide estimates of the heat emitted by the earth's surface, and the receipt of latent heat. The subsequent two columns (4 and 5) are estimates of terrestrial and solar radiation. The original figures in these tables have been converted into units of watts per square metre for achieving uniformity.

These observations were made in the mid sixties. They suggest net accumulation of heat west of 60°E, which is twice the value for the sector to the east of 60°E.

Table 10.2

Monsoon Heat Budget West 60°E Meridian (after Bunker, 1965)

Unit : Watts/m²

Pressure (mb)	*Sensible heat flux*	*Latent heat flux*	*Outgoing long-wave radiation*	*Absorbed short wave radiation*	*Net*
600 to 700	—6 5	10·5	—2·5	0·5	2
700 to 850	16·5	72·0	—26·5	10·0	72
850 to 1,000	54·5	124·0	—31·5	18·0	165
Total	64 5	206·5	—60·5	28·5	239

Table 10.3

Monsoon Heat Budget East of 60°E Meridian (after Bunker, 1965)

Unit : Watts/m²

600 to 700	8·5	14·5	—16·5	6·0	12·5
700 to 850	5·0	23·0	—22·5	12·0	17·5
850 to 1,000	5·0	83·5	—23·5	17·0	82
Total	18·5	121·0	—62·5	35·0	112

If we compare the two tables, we find that the flux of sensible heat in the western sector of the Arabian Sea (column 2 in Tables 10.2 and 10.3) is considerably greater than the corresponding value for the eastern sector. The greater intake of heat from the sea surface indicates cooler and drier air over the western sector of the Arabian Sea than over the eastern sector. This, as indicated earlier, is because of the Somali Current, upwelling and a low level inversion over west Arabian Sea.

It is also interesting to find that the net loss of heat by terrestrial radiation is about the same in both sectors of the Arabian Sea, although there is a difference in its distribution with height. The western air is generally dry, which leads to a small loss in terrestrial radiation from the top layer between 600 and 700 mb (Table 10.2). Similarly, the absorption of short-wave solar radiation by the eastern air (Table 10.3) is larger because of its greater cloudiness and haziness. There is need today of more observations of the type which we have described. In particular, we need to know how the heat reservoir over Pakistan, the Red Sea and

Arabia builds up in the course of the three months of the pre-monsoon period.

The Transport of Water Vapour During the Monsoon

Estimates of the transport of water vapour were first made by P.R. Pisharoty from India in 1965 after the International Indian Ocean Expedition. The observational data suggest that air over the equator is comparatively dry. This conclusion was based on the aerological observations over Gan Island (74°E, 0°S) which is a small island in the Indian Ocean slightly to the south of the equator. Upper air soundings from this island are available for a few years, and it has been observed that the moisture content of the air above this island is low compared to other stations along the west coast of India.

With a relatively dry air mass near the equator, it seems reasonable to infer that for the monsoon the main intake of water

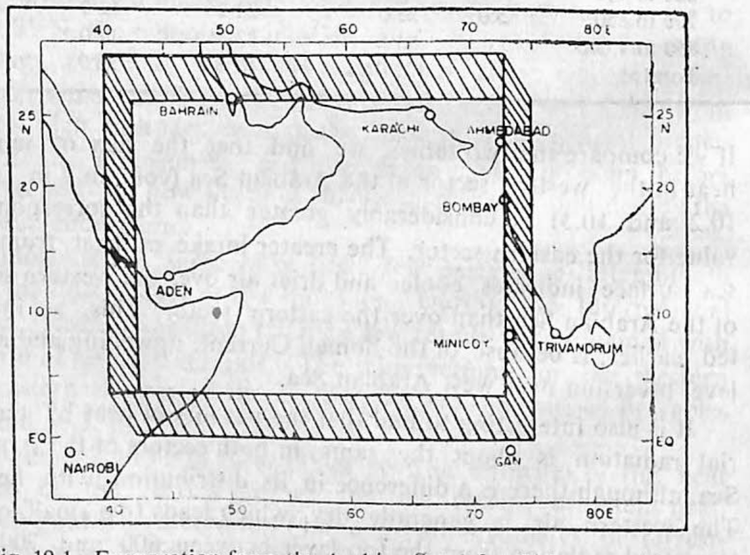

Fig. 10.1—Evaporation from the Arabian Sea (after Pisharoty, 1965).

is by evaporation from the Arabian Sea. Indeed, it has been estimated that the total transport of water vapour across the equator is only about a third of what is transported into the Indian peninsula across the west coast of India on a typical monsoon

day. Dr. Pisharoty made a quantitative estimate of the amount of water that could be picked up by evaporation from the Arabian Sea. He computed the flow of water vapour across the four lateral walls of a hypothetical rectangular box. This is shown in Fig. 10.1. The bottom of the box represents the sea surface and a rigid lid appears at 450 mb. This is the upper limit for which observations of water vapour content are available from our aerological soundings. Of the four lateral boundaries, the southern one appears along the equator, the western one along 42°E, the eastern along 75°E and the northern one was placed along 26°N. Computations reveal that the average net outflow from the four walls of the box was 3.4×10^{10} metric tons per day in July 1963 and about 1.2×10^{10} metric tons per day in July 1964. The reason for the large difference between the two years is rather vague. In view of the difficulties in making computation of this nature, we may treat these figures as representing orders of magnitude of the outflow of water vapour from the Arabian Sea.

During the Indo-Soviet Monsoon Experiment of 1973, four Soviet research vessels made extensive cruises over the Arabian Sea and the adjoining north Indian Ocean, and took six hourly upper air soundings. Upper air measurements of wind, temperature and dew point temperature were made for the first time. Using these observations, computations of the water vapour transport, for the active and weak monsoon phases were made by S.K. Ghosh and his collaborators in India. The computed values are shown in Table 10.4.

Table 10.4

Water vapour flux (10^{10} metric tons per day)

Boundary	*Monsoon*	*condition*
	Weak	*Active*
Western (50°E)	2.45	2.53
Southern (the equator)	3.23	2 93
Eastern (75°E)	7.12	12.50
Northern (latitude 20°N)	1.19	0.30

The above values indicate that the water vapour transport during an active phase of the monsoon across the west coast of India is more than four times that across the equator.

Secondly, the water vapour flux across the equator shows no significant difference between active and weak monsoon conditions. The large variations of water vapour flux across the west coast of India during the two contrasting phases of the monsoon are apparently not related to the transport from the southern hemisphere. This leads one to believe that fluctuations of rainfall over the west coast of India are more closely related to changes in monsoon circulation over the Arabian Sea.

These figures neglect the amount of water that is received by the atmosphere from daily precipitation over the Arabian Sea, as well as the upward flux through the rigid upper lid at 450 mb. The measurement of rainfall over a sea surface is an extremely difficult proposition, and no wholly satisfactory method has been yet devised. It is also difficult to estimate the water vapour content of the atmosphere above 450 mb because the moisture sensing element of the radiosonde becomes ineffective at this level. These estimates that have been made are, therefore, subject to an error of about 15%.

At present there are a number of uncertainties concerning the transport of water vapour which have not been fully resolved. In particular, when we consider evaporation from the sea surface we find that its rate depends strongly on the prevailing winds and the thermal structure of the atmosphere immediately above the water surface. When the wind strength is less than 7 m sec^{-1}, the surface is by and large calm, but at higher wind speeds the surface is rough. These features control the rate at which water vapour is lost by evaporation.

We also draw attention to the lack of accurate information on the distribution of rainfall over large ocean areas. Until we have accurate observations of precipitation over the oceans of the globe, it will be difficult to make an assessment of the accuracy of evaporation surveys. And, we need hardly emphasize the importance of an accurate knowledge of evaporation in the study of atmospheric energetics.

The necessity for the transport of water vapour in the atmosphere mainly arises from the existence of an excess of precipitation over evaporation in certain parts of the world, and a reversal of such conditions in other areas of the globe. Over a sufficiently long period of time, the amount of precipitable

water in the earth's atmosphere does not change appreciably. This means that the storage of water vapour in the earth's atmosphere is small and, as a consequence, large scale deficits and excesss of water vapour must be made up by the transport of water vapour through atmospheric circulations. With improvements in the network of aerological stations distributed over the northern hemisphere, it is now possible to make approximate estimates of the transport of water vapour on a global scale. In the table following, we present a survey of water vapour flux across different latitudes of the northern hemisphere.

Table 10.5

Water Vapour Flux Across Specific Latitudes for the Northern Hemisphere

Unit : 10^{10} metric tons per day

Lat.	*Winter*	*Summer*	*Annual*	*Lat.*	*Winter*	*Summer*	*Annual*
80	—0·09	—0·10	—0·10	40	4·63	4·87	4·72
70	0·55	0·41	0·47	30	3·29	1·82	2·51
60	1·94	2·11	2·04	20	—3·18	—1·59	—2·33
50	3·87	5·88	4·60	10	—12·32	1·66	—5·32
45	4·35	5·98	5·01	0	8·05	7·85	0·0

The flux is reckoned to be positive if it is directed northward, and negative if its direction is southward. The figures corresponding to 0, 10° N and 20°N for summer are of direct interest to us because they correspond to the period of the Indian monsoon. The northward transport is 1.6×10^{10} metric tons per day at 10°N in summer. This figure is about the same as Pisharoty's estimate of the flux across the equator.

These figures reveal a reversal of the transport of water vapour between winter and summer. It will be observed that the flux between the equator and 10°N is strongly negative during winter, but becomes positive in summer. When the summer and winter maps are compared, the seasonal reversal of direction is well marked over most parts of the Bay of Bengal, south China Sea, the equatorial portion of the western Pacific Ocean, over the Gulf of Guinea and the adjacent regions of Africa. This oscillation represents a net positive flow of moisture across the equator during summer and an opposite flow during winter. It is undou-

btedly a reflection of the monsoonal reversal of winds, and the well-known northward shift of the Intertropical convergence zone in summer.

Before we close our survey of water vapour transport, it is worth emphasising that a high value of water vapour flux does not necessarily ensure more precipitation. But, those regions which exhibit strong convergence in the flux of water vapour are usually associated with large amounts of precipitation. We see that during summer the northward flux of water vapour across the equator is 7.9×10^{10} metric tons per day. By the time the air travels to 10°N, the flux is reduced to about one-fifth of its value at the equator, at 20°N the direction of the flux is actually reversed. Thus, we can infer that the region between the equator and 20°N is one of considerable convergence, or net accumulation of water vapour during summer. It is no accident, then, to find that this is also a region of considerable precipitation.

The Monsoon and the General Circulation of the Atmosphere

The general circulation of the atmosphere has been the subject of considerable research by meteorologists of many generations. A large number of theories to explain the behaviour of winds is to be found in meteorological publications of the nineteenth and early twentieth century.

We shall not attempt a detailed survey of the subject here because our main concern is with the limited field of the Indian summer monsoon. But, we should like to refer to a review article by Professor Lorenz in 1967, to whose insight we owe many of the modern concepts of the general circulation.

As we have seen in the earlier chapters, an early attempt to explain the trade winds near the equator was due to Edmund Halley in 1686. He ascribed the existence of northeasterly trades

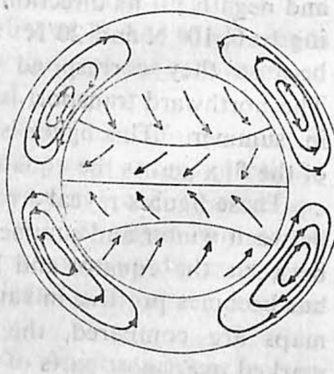

Fig. 10.2—Schematic representation of the general circulation of the atmosphere as envisioned by Hadley in 1735 (after Lorenz 1967).

to the northern side of the equator and southeasterly trades to the south of the equator as a tendency for air to converge towards the most strongly heated regions of the earth, namely, the equator.

Shortly after the time of Newton, Hadley, in 1735, published an account of trade winds which was an improvement on the earlier concept of Halley. In agreement with Halley, Hadley suggested that the distribution of solar heating would lead to rising motion in the equatorial regions and sinking motion near the poles. To compensate for the upper flow from the equator to the pole, Hadley envisaged a return flow from the pole to the equator at lower latitudes. A schematic representation of the general circulation, as envisaged by Hadley, is in Fig. 10.2.

The earth's surface moves most rapidly eastwards at the equator. Hadley, therefore, reasoned that if air were moving towards the equator from higher latitudes, it would arrive at lower latitudes moving westward relative to the earth. The air would have a tendency to conserve its absolute speed of motion. When it arrived at low latitudes it would, as a natural consequence, tend to fall behind the comparatively faster moving air at the equator. Air arriving at low latitudes would therefore appear to come from an easterly direction (Fig. 10.2). In fact, Hadley found that if air travelled over considerable distances, it would acquire a much greater westward velocity than was ever observed. The fact that such high velocities were never observed in the real atmosphere was explained as a consequence of the frictional drag of the earth's surface on the atmosphere.

The effect of the earth's rotation, or the Coriolis force, on the large scale movement of air was pointed out by William Ferrel in 1856. The Coriolis force tends to deflect air to the right in the northern hemisphere. The magnitude of this force also increases with latitude. Consequently, as the air ascends over the equator and spreads out polewards, it is deflected to the right and soon acquires an eastward component of motion. The eastward component increases sharply as the air moves into higher latitudes. Gradually, therefore, the air becomes a westerly wind as it begins to sink at higher latitudes. At lower levels, we have a return current from the pole to the equator. The effect of the earth's rotation decreases progressively as the air moves towards the

equator in the return current. The westerlies, therefore, weaken and change into easterlies as they approach the equatorial regions.

Ferrel drew our attention to the fact that Hadley's model did not consider the radiative loss of heat from the upper layers of the atmosphere. Towards the beginning of the chapter, we mentioned that as we proceed to higher latitudes the loss of energy by outgoing long wave radiation exceeds incoming radiant energy from the sun. The loss is equivalent to a cooling of the atmosphere by about 1° to 2°C per day. As a consequence of radiative cooling, the equatorial air would have been sufficiently cooled by the time it reaches about 30°N, to sink and spread out horizontally at the ground. A part of the subsiding air may be expected to return towards the equator, while another part would tend to spread out towards the polar regions.

This line of reasoning enables us to picture the general circulation of the atmosphere as being made up of three cells on the meridional plane. We have a cell between the equator and about 30°N, in which there are easterlies (trade winds) at lower levels and westerly winds aloft. There is then a middle cell between 30°N and 60°N in which there are westerly winds at the earth's surface and easterly winds aloft. Finally, we have a polar cell between 60°N and the pole, where the surface winds are easterly and the upper winds are from a westerly direction. The model is shown schematically in Fig. 10.3.

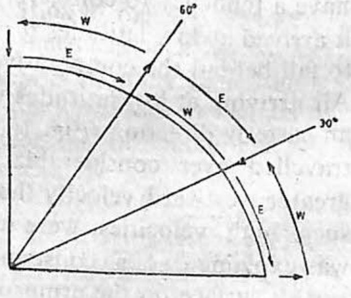

Fig. 10.3—A Cellular model of the general circulation.

Later studies have emphasised one major deficiency of the above model. It predicts easterly winds at the upper levels in mid-latitudes. But, as we can see from observational evidence, in the latitude belt between 30°N and 60°N, the westerly winds extend throughout the entire depth of the troposphere and a part of the lower stratosphere. Moreover, it is known that between 300 and 700 mb the westerly winds often reach speeds exceeding 60 knots in the form of a jet stream. In an attempt to get over this difficulty, the eminent Swedish Meteorologist, C.G. Rossby, suggested

that upper easterly winds in the middle cell were wiped out by the frictional drag exerted by upper westerly winds to the north and south of the middle cell. There are problems of this nature in the general circulation of the atmosphere whose answers are not yet known.

Tropical meteorologists are, however, mostly concerned with the equatorial cell in which we have trade winds at lower levels and westerlies aloft. In the last two decades, meteorologists have increasingly realised that, as in other branches of physical sciences, it is best to search for conservation principles while studying the large scale movements of air. To physicists, one of the most familiar conservation principles is that of mass. With the help of this principle we can assert that there can be no physical transformation in nature in which mass is created out of nothing; nor can we envisage a change in which a certain amount of mass is destroyed for good.

In a similar strain, it may be reasoned that when air moves across considerable distances from a higher to a lower latitude, the characteristic which it conserves is not its absolute speed of motion, but its angular momentum. In physics, we refer to the product of the mass and the velocity of a moving body as its momentum. For bodies which rotate round an axis, such as spinning top, or an envelope of air round a rotating earth, it is more convenient to consider angular momentum. The angular momentum of a body is the product of its momentum and its distance from the axis of rotation. In view of its larger distance from the earth's axis of rotation, we readily see that air at low latitudes has greater angular momentum than air at higher latitudes.

The angular rotation of the earth is from the west to the east. In low latitudes, the prevailing winds are easterlies; consequently, the atmospheric winds oppose the movement of the earth. This is equivalent to stating that, at low latitudes, the earth loses angular momentum to the atmosphere. Conversely, in middle latitudes, the prevailing winds support the movement of the earth. Consequently, the earth gains angular momentum in these regions at the expense of the atmosphere. There is also a slight loss of angular momentum in the polar regions, where surface easterlies prevail, but the quantity of angular momentum involved here is very small because of the small distance from the earth's axis of

rotation. We have, therefore, a situation in which the earth is continuously losing angular momentum in low latitudes and extracting angular momentum from the atmosphere in mid-latitudes.

Now, one could legitimately argue that the earth's speed of rotation has shown no appreciable change with time; nor have there been any systematic changes in the speed of different zonal wind belts. There must be a balance, then, between what the atmosphere gains in angular momentum at low latitudes and what it loses around the mid-latitude belt. To maintain the state of balance the surplus angular momentum of the tropical atmosphere must be transported to the belt of westerlies.

A peculiar situation prevails during the Indian summer monsoon. During this period the prevailing winds over the Indian peninsula and adjoining areas of the Indian Ocean are westerly at low levels with easterly winds aloft. The monsoon westerlies, therefore, lose momentum to the earth in contrast to the easterly trade winds. In recent years, Indian meteorologists have tried to compute the flux of angular momentum during the southwest monsoon to see if the balance requirements are fulfilled. The upper winds over Singapore suggest that there is a large transport of angular momentum northwards across the equator to partially compensate for the loss of angular momentum in the regime of the westerly monsoon winds. A more recent study of the balance of angular momentum suggests that most of the angular memen-tum that is fed into the Indian region is through meridional motion. By meridional motion we mean the northward movement of air along a meridian.

When we consider the flux of angular momentum over limited regions of the earth, it is convenient to consider a rectangular region bounded by two parallels of latitude and by two meridians. Calculations have been made for a region bounded on the west and the east by the meridians 50°E and 100°E, and along the north and south by the equator and 20°N. This region includes most of the Indian Peninsula, Burma, Indonesia and Gulf of Aden. The computations reveal the largest contribution to the flux of angular momentum in the lower troposphere is by meridional motion. This is 7×10^{25} gm cm^2 sec^{-2}. This flux is of sufficient

magnitude to maintain a westerly belt of zonal winds against the frictional drag of the earth's surface.

Calculations of this nature lend some support to our concept of a northward shift of the southern hemisphere trade wind system across the equator during the monsoon months. These computations suggest that the northward shift feeds angular momentum of the right order of magnitude into the monsoon circulation so that it can survive against the frictional drag of the earth's surface.

CHAPTER XI

MOUNTAINS AND MONSOONS

MANY FACETS OF Indian climate are influenced by the Himalayas. This massive barrier is shaped like an ellipse which extends beyond a thousand kilometres both along the east-west and north-south directions. It contains some of the largest peaks and the deepest valleys of the world. The major river systems of India, namely, the Brahmaputra, the Ganges and the Indus have their origin in the Himalayas. When we speak of climate, it is necessary to distinguish between weather on different scales.

In the first category, one might consider the genre of local winds, that are generated by individual peaks and valleys. This is motion on a small scale influenced by the size of a peak and a valley, that is, about a hundred and fifty kilometres in length and about fifty kilometres in width.

But motions also occur on a larger scale. Here one is concerned with the impact of the Himalayas on the large scale weather systems of Indian climate, namely, the monsoons and jet streams or the passage of disturbances from the west to the east across the Himalayas.

There is a sharp difference between the time scales of motion in these two categories. The first occurs on the scale of a day. But the second category is on a larger time scale, because it occurs over several days. The important point is that climate is an integrated effect of motion on different scales of time and space. When we refer to climate generated by individual peaks and valleys for example, we will be largely concerned with mountain and valley winds that could occur at any time of the year. But, when we consider the larger scale of motion, such as,

the distribution of rainfall, it will be apposite to refer to particular months or seasons.

It will be hardly correct to assume that Himalayan climate was entirely dominated by winds. Winds are generated by interactions between a wide variety of meteorological variables, such as, the pattern of pressure around large peaks and the vertical profiles of temperature. The distribution of clouds and the radiation received from the sun are also important because of the high reflective properties of ice and snow.

Much of our difficulties arise from lack of data. It is difficult, if not impossible, to set up conventional data platforms on mountains. Fortunately, this situation is now beginning to improve with the advent of weather satellites. Even now we are beginning to get better information on snow cover by intercepting the pictures transmitted by satellites launched by India, the USA and USSR. Such pictures can now be received with very high resolution.

Seasonal features

As over the plains of India, the climate of the Himalayas is largely seasonal in nature. The principal seasons are :

(*i*) Pre-monsoon (March-May)
(*ii*) Monsoon (June-September)
(*iii*) Post-monsoon (October-November)
(*iv*) Winter (December-February)

There are active and comparatively quiet periods within each season. Such periods need not of necessity coincide with similar conditions over the plains of India. Thus, what would be a comparative lull in rainfall over the plains of north and central India could well be a period of "active" or heavy rain over the eastern Himalayas. There is an opposition in phase between the eastern and western sectors of the Himalayas during the monsoon season.

The seasonal character of the climate is often reflected in the strength of upper winds over the Himalayas. During winter and the post-monsoon period, strong westerly winds are a characteristic feature of the mountains, but during the monsoon there is an abrupt northward movement of the belt of strong winds.

By and large, the characteristic features of each season make up the climate of the Himalayas. Let us now proceed to details.

Motion on small scale—Mountain and valley winds (50-100 km)

Barriers across the path of a large stream of air create situations of much meteorological interest. The winds adjacent to a mountain slope, or the stagnant air trapped in a valley, become heated during day time by the radiation received from the sun. This generates strong ascending currents of warm air up the valley and along the mountain slopes. The day time updraft of warm air is frequently indicated by the presence of large cumulus clouds near mountain peaks. With the setting of the sun the situation is reversed. The slopes of the mountain cool very rapidly, thereby chilling the air layers immediately above them. The cooler and heavier air then begins to slide down the mountain slopes into the valleys.

The ascending current during day time is known as anabatic wind, while the cooler descending currents of night are named katabatic winds. They are observed in the vicinity of individual peaks and adjoining valleys.

Mountain winds of this nature are well known in many parts of the world. Winds, such as, the "bora" off the northeast coast of the Adriatic Sea and the "mistral" in southern France are good examples of mountain winds. The narrow northwestern coastline of the Adriatic Sea is backed by a high plateau where the air becomes very cold. By reason of its larger density, this air slips down the plateau as a cold, gusty wind whose temperature is much colder than the air along the coast below.

An expedition to Khumbu Himal region of the Nepal Himalayas by a team of scientists from Japan recorded interesting observations of upslope winds during day, and down-slope winds at night in the vicinity of Lhajung (27.53°N, 86.50°E), which is at an altitude of 4420 m. It is located on a moraine terrace elevated by 200 m from the floor of a U-shaped valley. Reports provide evidence of a well marked diurnal variation in the rainfall pattern of this valley, which could be attributed to upslope winds during day and downslope winds at night.

We have very little by way of documented data on the energy exchanged between mountain winds which are confined to

individual peaks and adjoining valleys, and the circulation of the free air above the peaks. Such a study would be interesting because in the Himalayas, there are several major valleys running at right angles to the main mountain ranges. Mountain winds are known to have an effect on the shape and the type of trees that flourish on mountain slopes. There is evidence to show the deformation of coniferous trees in the Kali Gandak valley in Nepal as a consequence of frequent exposure to valley winds.

Large scale motion (1000-5000 km)

The impact of the Himalayas on large scale atmospheric circulation may be considered under :

— The Tibetan anticyclone and monsoons
— Movements of a west to east jet stream
— Eastward passage of disturbances from the Caspian Sea to the east to Tibet (western disturbances)

We will consider these features in turn.

Tibetan anticyclone and monsoons

In the summer months of April and May, the semi-arid regions of northwest India receive more solar radiation than the adjoining land areas. This warms up the earth's surface. And, it is easier to warm the air above land by solar insolation than it is to warm the sea or the oceanic regions to the south of India, because water has greater heat capacity. The effect of differential heating is to set up a large current of air from the Indian Ocean, south of the equator, towards India. This is the summer monsoon.

In winter the land cools much faster than the sea. Consequently, there is a reverse sweep of air from Siberia to the Indian Ocean. This generates the winter monsoon.

Reports indicate that most of the precipitation recorded over the high altitudes of the Nepal Himalayas falls during the monsoon months from June to September. The accumulation of ice on the glaciers of this region depends on the quantity, and the variation of precipitation during these months.

Short period variations of rainfall over the eastern Himalayas are also related to the periodic movements of the "monsoon trough". The monsoon trough is an elongated low pressure system over the plains of north India. When the axis of the

monsoon trough lies close to the Himalayan foothills, there is an increase in rainfall over the eastern sector of the Himalayas. As many of the rivers that flow through India have their origin over the eastern Himalayas, this situation often leads to floods over the plains of Assam, north Bihar and sub-Himalayan West Bengal. The duration of this situation is usually about 5 days, but breaks that occur towards the end of the monsoon in August and early September are of longer duration.

Indeed it has been observed that there is some periodicity in the rainfall data of Nepal. There appears to be a predominant periodicity of 10 days and a secondary cycle of 5 days. The periodicity of 10 days occurs simultaneously over all stations, but the cycle of 5 days represents the westward passage of disturbances with a speed of about 500 km per day.

The westerly jet stream

The jet stream is a narrow current of very fast moving air. This feature of the atmospheric circulation is very well marked over north India and the Himalayas in winter. The core of the west to east jet stream is usually located around 10 km and winds exceeding 60 knots are not uncommon. There are small north to south movements in the core of the westerly jet, especially with the passage of eastward moving disturbances in winter.

A peculiar feature of the jet stream is its rapid movement to the north of the Himalayas just before the onset of the monsoon. In fact, throughout the monsoon season, the jet stream remains much to the north of the Himalayas. But, with the advent of winter, it again reappears.

The Himalayan barrier is often effective in splitting the westerly jet stream into two branches. In winter, it is often observed that one branch is located to the north of the Himalayas, while another branch is situated to the south. The two branches join together again over northern China and Japan.

The base of the jet stream is a little higher (9 km) than the highest Himalayan peaks, but even so its effect is felt in the major valleys and peaks of the Himalayas. We have little data so far to indicate the type of energy exchanges that take place between the jet stream on the one hand, and mountain winds (upslope and downslope) on the other.

Western Disturbances

During the post-monsoon period (October and November) and the winter season (December to February) weather over the Himalayas is determined by the eastward passage of low pressure systems. These systems have their origin as cut-off lows over the region adjoining the Caspian Sea. They move from the west to the east and are known as Western Disturbances to meteorologists in India. Their frequency varies from year to year but, on an average, one expects 3-5 disturbances per month.

Precipitation, often in the form of snow, is observed ahead of each western disturbance. Most western disturbances do not have the frontal structure that is typical of mid-latitude low pressure systems. On some occasions an occluded front may be discerned, but the temperature contrast between the different sectors of these systems is not well defined. Many disturbances are split into two or more secondary systems by interaction with the mountain ranges on the western sector of the Himalayas. These secondary systems, along with the main disturbance, provide the plains and foothills of north India with winter rainfall.

Western disturbances are responsible for a significant difference in the rainfall distribution over the eastern and western sectors of the Himalayas. The eastern sector receives most of its rainfall during the monsoon months of June to September. The western ranges, namely, the Pamir-Karakoram ranges, receive very little by way of monsoon rain. On the other hand, the western half of the Himalayas receives more precipitation in winter from western disturbances.

On some occasions, there is a revival in the intensity of a western disturbance as it moves over Tibet. This is explained by the fact that even in winter the southeastern parts of Tibet act as a weak heat source. Consequently, even at 500 mb which is roughly the elevation of Tibet, the air is warmer than the ambient air at the same level over regions adjoining Tibet. Thus, if the cold air associated with a western disturbance breaks out over southern Tibet, a large horizontal temperature gradient is built up around the southern slopes of the plateau of Tibet. This temperature contrast provides a mechanism for greater input of energy into eastward moving low pressure systems.

The interaction between western disturbances and the westerly

jet stream has not yet been quantified. We often see a weakening in the strength of the westerly jet with the advent of a western disturbance, with a strong revival in intensity to the rear of the disturbance. A similar feature is observed on the northern branch of the jet, whenever we have large amplitude troughs from the mid-latitudes moving eastwards across the northern fringes of the Himalayas.

Special meteorological variables :
Radiation

Atmospheric circulations derive their energy from the sun. The important quantity which we need to measure is the difference between (*i*) the incoming radiation from the sun in short wave lengths (0.2 to 4 microns), and (*ii*) the outgoing long wave radiation emitted by the earth's surface (4-80 microns). Unfortunately, this is very difficult to measure because it represents a small difference between two large quantities. An added difficulty lies in the fact that both the downward and upward radiation are modified by (*i*) reflection by earth's surface and by clouds, (*ii*) absorption by atmospheric constituents, such as, water vapour, carbon dioxide and ozone and (*iii*) scattering by atmospheric aerosols. None of these changes have been well documented over the Himalayas. Some data are now beginning to emerge from observations recorded by mountaineering expeditions, and from recent observations with the help of weather satellites.

The direct solar radiation has been observed to be around 1150 watts per square metre. A similar value of 1200 watts per square metre was observed at Lhajung in Nepal during the month of May. Subsequently, this begins decreasing in June with the beginning of the rainy season. In January, the direct solar radiation is reduced to around 500 watts per square metre. These are approximate estimates because there must be wide variations depending on the location of the observation point. Usually, the southern slopes tend to receive more radiation than those facing northwards.

Generally, in the monsoon season not only the direct solar radiation but the downward reflected radiation from clouds increase, because of increasing cloudiness. The result is a surplus of radiation over the Himalayas in summer. In winter, the

situation is reversed because of decreasing cloudiness. It has been estimated that the nocturnal radiation in December is about 20 per cent more than the net radiation, which leads to a small deficit in balance.

Atmospheric Aerosols

Apart from their role in the radiation balance of the earth-atmosphere system, it is important to study the nature of aerosols over the Himalayas, because the mountains are located far away from industrial regions. Aerosols over the Himalayas are a measure of the background pollution of the free atmosphere.

Observations have been made at Shorong in eastern Nepal during the monsoon and over Muktinath in the middle western sector of Nepal in the winter season (November). These observations reveal that most of the samples that were collected at these two locations were sulphate compounds, mainly Ammonium Sulphate.

Most particles were in the range 0·05 to 0.2 microns ; although a few particles of radius greater than 0.2 microns were found. The concentration of smaller particles varies from 9 to 170 per cubic centimetre. These concentrations suggest considerable contamination of the atmosphere, perhaps by human activity in nearby villages.

Special features of precipitation

A characteristic feature of rainfall is that most of it occurs at night. Why this should be the case is unclear at present, but it could be an outcome of diurnal variations in the strength of local winds.

But all phases of precipitation have an important effect on the hydrology of Himalayan glaciers. Precipitation occurs either in the liquid phase (rain or drizzle) when the temperature is greater than 0°C, or in the solid phase (snow or ice pellets) at temperatures below 0°C. In the temperature range 0°C to —3°C, the precipitation could be a mixture of the liquid and solid phases.

This feature of the precipitation is important for determining the mass balance of glaciers. The snow accumulates over the high peaks of Himalayas during the post-monsoon and winter season. The accumulation is maximum during the winter when the snow line comes down to about 1500 metres in the western Himalayas

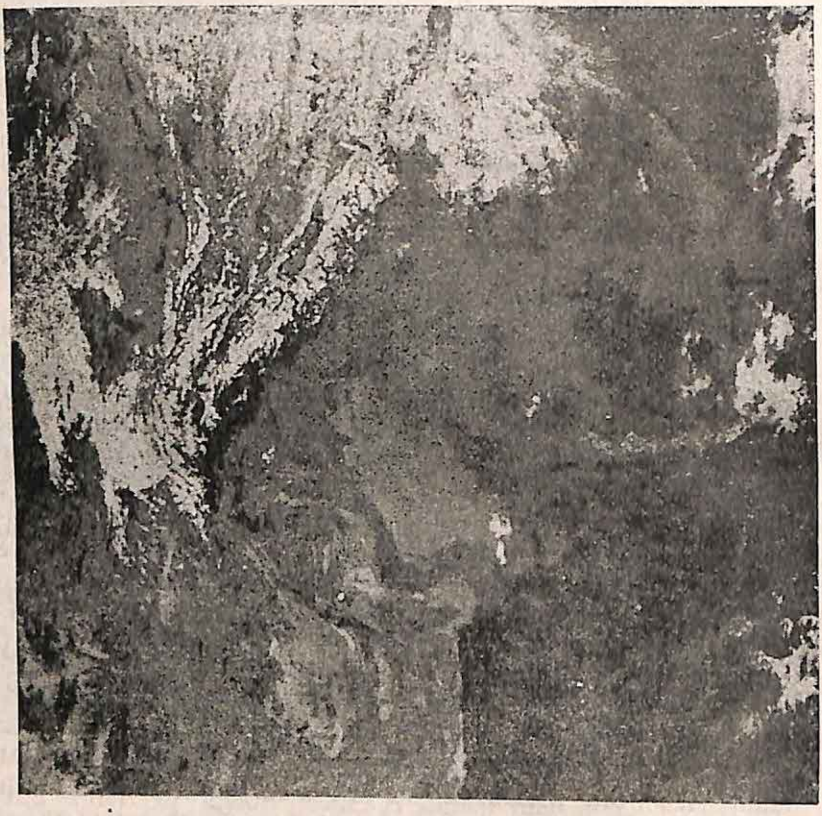

Fig. 11.1—Photograph of May 7, 1982 from Indian Geostationary Satellite (INSAT—IA) depicting dendritic pattern of snow cover over Himalayas.

and to about 3000 metres in the eastern Himalayas. The snow melt starts in summer and contributes significantly to the discharge from the major rivers of India. During the monsoon, only the perennial snow line remains over the peaks above 6000 metres. A satellite picture of snow covered Himalayas is given in Fig. 11.1.

Ecological Imbalances

An interesting idea which is attracting much research in recent years, is concerned with measures to accelerate snow melt and to reduce the loss of snow by evaporation. Two techniques have been

used for this purpose. The first is to spread a thin film of alcohol on a melting snow surface to suppress evaporation. The other experiment deals with spreading lamp black to accelerate the ablation of snow. Lamp black is a black body which absorbs radiation from the sun and decreases the reflectivity of snow. There are three basic questions which still remain unanswered : (*i*) will an alcohol film spread on a melting snow surface remain sufficiently long to bring about reduction in vapourisation, (*ii*) will these materials remain on the surface for a sufficiently long time and (*iii*) what magnitude of water might be obtained by this process ?

Firm answers to these questions have not yet been obtained but the experiments appear to be sufficiently interesting to warrant further research. The danger to which we referred earlier, namely, indiscriminate cutting down of forests could also have an adverse effect on snow cover.

We have presented in this chapter some of the interesting features of the climate of the Himalayas. There are many features which we still do not understand. Our knowledge on these features will improve only when we acquire data and better monitoring facilities. But there is one aspect which needs caution. There are sufficient indications of human activity having an adverse impact on climate. Removal of trees and vegetation is an example. This is specially relevant for the Himalayas because indiscriminate cutting down of forests could, under certain conditions, decrease the rainfall of this region. There are dangers of ecological imbalances being set in motion, and it is not too early to start thinking of protective measures against them.

The Flow of Air over the Western Ghats

Apart from the Himalayas, there is another mountain barrier which is important for the Arabian Sea branch of the monsoon. We refer to the Western Ghats, which run roughly parallel to the western coastline of India. The Western Ghats represent a series of mountains that rise abruptly off the coastal plains to an altitude of 1 to 2 km. The moist air currents which approach the Western Ghats from the west are forced to ascend the mountains. In this process, they shed their moisture in the form of frequent and heavy rain over the Western Ghats.

After they have surmounted the Ghats, the monsoon winds advance into the Deccan plateau, Madhya Pradesh and thence to the Bay of Bengal. But, as they have already shed their moisture, it is not surprising to find a strong contrast between rainfall on the west coast and regions to the lee of the Ghats.

When an air mass strikes a vertical obstacle it is forced to ascend. Often it may happen that the air does not have sufficient energy to surmount the obstacle. On such occasions, winds tend to go round the barrier rather than over it. When we consider very high barriers having the dimensions of the Himalayan massif, physical intuition tells us that a large part of the air would have a tendency to go round the mountains rather than climb over them. But let us consider a barrier of the size of the Rocky mountains or the Western Ghats. The striking contrast in rainfall on both flanks of the Rockies or the Western Ghats suggests that the air climbs over the mountain.

How can we be sure if an approaching air mass has sufficient energy to climb over an obstacle in its path ? There is a theorem on the motion of fluids which helps us to throw some light on this problem. This theorem is named after an eminent mathematician, Daniel Bernoulli. Bernoulli's theorem tells us that the sum of the potential and kinetic energies of a fluid remains constant during its motion. In the parlance of physicists, we define potential energy as a type of energy which exists in form, as distinct from kinetic energy, which represents a measure of the energy of motion. If we consider a column of air whose centre of gravity is located at a height z above the earth's surface, then its potential energy is $g \times z$, where g is the acceleration due to gravity.

Now, it can be argued that as long as a column of air is ascending a mountain, its centre of gravity is being continuously raised. This implies an increase in the potential energy of the rising air. But, as we can see from Bernoulli's theorem, the sum of the potential and the kinetic energy is invariant; consequently, an increase in the potential energy of ascending air means that its kinetic energy is being proportionately decreased.

If we know the elevation of the barrier, it is fairly straightforward to compute the increase in potential energy as the air climbs to the top. In addition, if we also know the speed at

which the air strikes the barrier, it is possible to find out if it does indeed possess sufficient kinetic energy to enable it to climb over the barrier.

With the help of this line of reasoning it is possible to show that a westerly stream of air, such as the monsoon, would not be able to climb over the Ghats unless it was fed with some other source of energy. The suggestion has been made that the additional energy needed for this purpose could be provided by the heat of condensation, which is released as the air loses its moisture during its ascent to the top of the Ghats.

Confirmation of this hypothesis has been provided by an investigation by R. P. Sarker on the flow of monsoon air over the Western Ghats. If we represent the Western Ghats by a barrier of approximately the same shape and elevation, it is possible to compute, on theoretical grounds, the rate at which air would be forced to ascend. Now, the moisture content of the monsoon air to the west of the Ghats may be derived from aerological soundings. Consequently, if we insert reasonable values of the moisture content, it is relatively straightforward to compute the rate at which the air would lose its moisture. If it be assumed, as a first approximation, that all the moisture that is shed by the air reaches the ground as rain, we have a method of evaluating what the rainfall pattern should look like. It is important to note that a comparison between the theoretical profile and observed rainfall over the Ghats would have more than purely academic interest. The comparison would tell us, for example, how much of the observed rainfall could be attributed to topography alone.

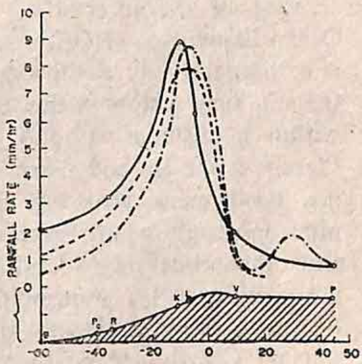

Fig. 11.2—Observed and computed rainfall over the Western Ghats (after Sarker). Full lines represent observed rainfall and dotted lines are for calculated values by different models.

Investigations that have been made so far suggest that about sixty per cent of the observed rain could be attributed to forced lifting of the moisture laden

monsoon air. On a few occasions, as much as eighty per cent of the observed rain was entirely due to the Western Ghats forcing the air to rise. In Fig. 11.2 we reproduce a diagram to illustrate the comparison between rainfall that was actually observed and the pattern that was theoretically computed. The comparison is for a typical monsoon day.

An important point in this diagram is the striking similarity between the calculated and observed maximum rainfall. In fact, computations suggest good agreement on most occasions between the maximum rainfall predicted by theory and what was actually observed.

On a day of strong monsoon conditions, computations suggest that the peak rainfall should occur at a distance of 10-12 km away to the west of the mountain crest. Under weak monsoon conditions, the maximum rainfall would be located about 25 km away from the crest of the mountain.

There is a sharp decrease in rainfall as we proceed away from the mountain crest to the lee of the Ghats. Unlike the Rockies however, there does not appear to be any strong contrast in temperature between the windward and leeward side of the Western Ghats. Certainly, the air to the lee is warmer, but not markedly warmer than the monsoon air on the windward side.

One of the interesting results of the International Indian Ocean Expedition (IIOE) was the observation that the monsoon is a comparatively shallow current over the Indian Ocean and the southern sectors of the Arabian Sea. But, as soon as it comes within a distance of about 200 km from the Indian coastline, there is a sudden and sharp increase in its depth. A suggestion has been made that the sudden increase in the depth of the monsoon could be attributed to the presence of the Western Ghats. But, theoretical considerations do not quite support this point of view. When the monsoon encounters the Western Ghats, it begins to rise at a distance of about 50 km from the crest of the Ghats. Clearly, then, there must be other forces which add buoyancy to the monsoon current as it approaches the Indian coastline.

One such factor may be the added instability of the air. In an earlier chapter, we saw that the distribution of density in the atmosphere i often favourable for large vertical displacements

of air. When this happens, the atmosphere is said to be unstable. For saturated air this implies a lapse rate of temperature in excess of 6°C per km. Instability in the atmosphere can make its presence felt in many ways. Sometimes, the criterion of lapse rate would suggest that the air is not really unstable. But, if the same air was given a small amount of lift by some external agency, the accompanying changes in lapse rate may make it unstable. We do not yet have sufficient aerological soundings to indicate whether the monsoon air does in fact possess this type of instability as it approaches India; but upper air soundings at Gan Island (0°41′S, 03°09′E), near the equator tend to suggest that the air is relatively stable near the equator. On the other hand, the fact that theory is only able to account for about sixty per cent of the observed rainfall over the Ghats suggests that the monsoon air does possess a certain amount of instability. It is likely that in the next few years, when we have more upper air soundings within the monsoon current, we will improve our understanding of this aspect of the problem.

CHAPTER XII

LONG RANGE FORECASTS OF MONSOON RAIN

WE HAVE OFTEN STRESSED the importance of rainfall for Indian agriculture. In view of its importance, meteorologists in India have tried for many years to predict the season's rainfall in advance. As early as 1880, the need for a seasonal forecast was one of the important recommendations by a Famine Commission for India.

A pioneer in this field was Sir Gilbert Walker, who tried to make seasonal predictions on the basis of associations between monsoon rain and preceding events in different parts of the globe. But before we proceed to details, it will be appropriate to explain a few statistical concepts that are useful when we deal with large volumes of data.

Correlation Coefficients

The coefficient of correlation is a statistician's device for finding out the relationship or association between two quantities that may appear at first sight to be unrelated. Let us consider, for example, the heights and weights of a group of twenty people. We could represent the data of every individual in this group on a graph, with heights set out horizontally and weights set out vertically. The points on the graph might well show a distribuion of points scattered over a wide area. It may immediately be apparent from the scatter of points that there is no obvious relation between an individual's weight and his height. But, it is equally possible that the scatter of points suggests a relation does exist between heights and weights, only we are not sure what the relation

is. The correlation coefficient is simply a measure of the relation between height and weight revealed by the data at our disposal. It is expressed as a number whose value varies between 0 and 1. If the correlation coefficient is 0, it implies that there is absolutely no relation between the variables in which we are interested. On the other hand, a correlation coefficient of 1 implies a perfect relation between the two. For example, if we have a correlation coefficient of 1, then all points representing individual heights and weights would lie on a single straight line.

The correlation coefficient by itself is not always an infallible guide. Sometimes, we come across spurious values of the correlation coefficient. For example, almost all meteorological variables exhibit some amount of correlation. But, in most cases the correlation coefficient is extremely low and, in general, stringent statistical tests are needed before we can be sure of the predictive value of a correlation coefficient.

Standard Deviation, Variance and Coefficient of Variation

Let us revert to the problem of representing the heights and weights of twenty people. An average or a mean value of the height or the weight will give us some information about the group; but often it is necessary to know how the observations are scattered around their average. For this purpose different parameters have been devised to measure the dispersion within a set of observations.

We illustrate this by using a simple example. Assume that a set observations consists of the first nine natural numbers 1, 2, 3, 4, 5, 6, 7, 8 and 9.

The average or mean value is simply the sum of all values divided by the number of observations. In this case the average or mean is 5.

An useful measure of the scatter of points is the Standard Deviation. If we compute the squares of the deviations of the observations from their average value, and evaluate the square root of the mean square deviation, we obtain the Standard Deviation. Consider, for example, the first nine natural numbers. Their average value is 5. The deviation of each number from the average, and the square of the deviation may be set out in the form of a table :

Table 12.1
Standard Deviation

X	X—5	(X—5)²
1	—4	16
2	—3	9
3	—2	4
4	—1	1
5	0	0
6	1	1
7	2	4
8	3	9
9	4	16

Total : 60

Standard Deviation $=(60/9)^{1/2}=2{\cdot}58$

On some occasions, it is preferable to work in terms of the Variance. The Variance is just the square of the Standard Deviation. In the example just considered, the Variance is $(2.58)^2 = 6.67$.

If we want to compare the scatter of different sets of observations about their means, it is useful to express the scatter in a form that does not depend on the size of the sample. For example, cats and dogs may be equally variable in length, but this may not be revealed by simply giving the Standard Deviation for one group of cats and another group of dogs. The Coefficient of Variation is found by expressing the Standard Deviation as a percentage of the mean value. We have,

$$\text{Coefficient of Variation} = V = \left(100 \times \frac{\text{Standard Deviation}}{\text{Mean}}\right)$$

Let us compare the relative variability of the following two sets of observations :

(*i*) Mean $=20$ (*ii*) Mean $=100$

Standard Deviation $= 2$ Standard Deviation $= 5$

For (*i*), $V=(100\times 2)\div 20=10$

For (*ii*), $V=(100\times 5)\div 100=5$

The observations in (*i*) are thus twice as scattered relative to their mean compared to those in (*ii*).

Coefficient of Variation of Monsoon Rainfall

The Coefficient of Variation of monsoon rainfall reveals interesting features. While discussing the climatology of the monsoon

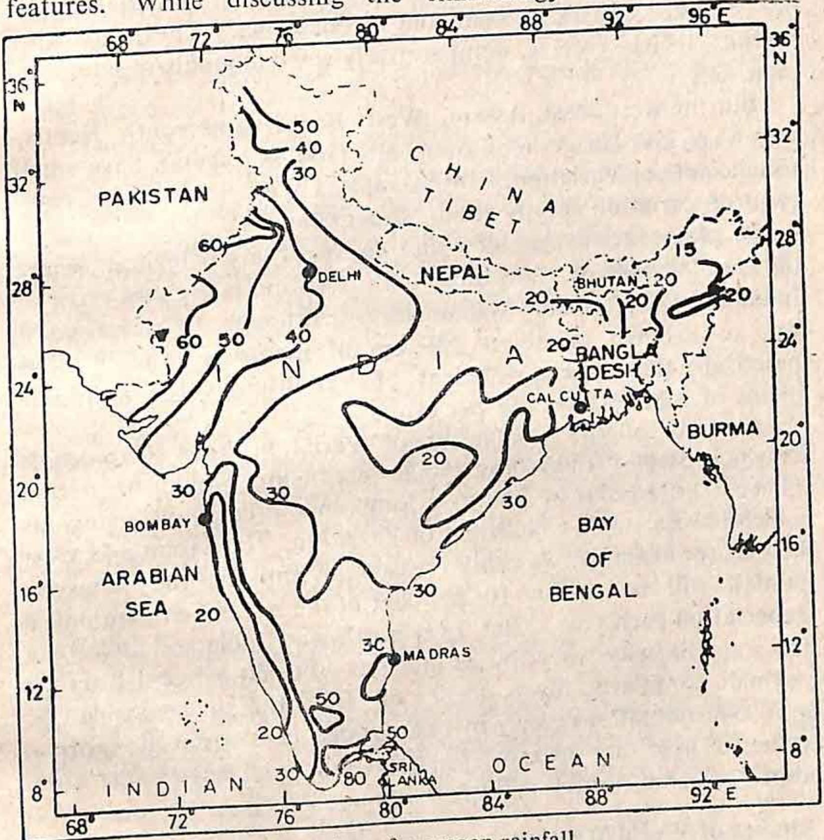

Fig. 12.1—Coefficient of variation of monsoon rainfall.
Based upon Survey of India map with the permission of the Surveyor General of India. © Government of India Copyright, 1988. The territorial waters of India extend into the sea to a distance of twelve nautical miles measured from the appropriate base line. Responsibility for the correctness of internal details shown on the map rests with the publisher.

(Chapter IV), we stressed the fact that wherever the rainfall was least, it was highly variable. This is well illustrated by the Coefficient of Variation of monsoon rain for different meteorological divisions of India. In figure 12.1 we present the values of the Coefficient of Variation for monsoon rainfall.

The Coefficient of Variation is highest in west Rajasthan (49 per cent) where the normal monsoon rainfall is only about 35 cm. The other regions which have a high Coefficient of Variation (30 per cent) are Gujarat, Punjab and Rayalaseema. These are also regions which receive comparatively small amounts of monsoon rain.

But the west coast, Assam, West Bengal and south Kerala, which receive large monsoon rain (100 to 250 cm), have small Coefficients of Variation. In Assam and West Bengal the Coefficient of Variation is only about 10 per cent.

In general terms, we infer that the variability is highest where the rainfall is least. An important exception is the west coast of India and the plains of Maharashtra. Although the west coast has an average of about 250 cm of monsoon rainfall, it has practically the same Coefficient of Variation as the adjoining plains of Maharashtra (18 per cent).

A study of the Coefficient of Variation helps us to decide in which parts of India long range prediction would be useful. There is little point in making a long range prediction for regions which have a small Coefficient of Variation. In Assam and West Bengal, for example, we can be reasonably sure that the monsoon rainfall will be within 10 per cent of the normal climatological expectation each year. But, over north-west India and Rajasthan the rainfall may vary by as much as 30 to 50 per cent from the normal. Similarly, over most of the Indian peninsula the monsoon rainfall may vary by over 20 per cent from its normal value. These are clearly the areas where advance information about seasonal rainfall will be useful.

Studies of World Weather

In his early work on the subject, Sir Gilbert Walker felt that the overall effect of energy received from the sun was to set up impulses or natural oscillations in the atmosphere. As a measure of solar energy he considered the number of sunspots to be an important parameter. In his earlier papers we find a wealth of correlation coefficients between sunspots and other meteorological variables, such as, the rainfall, pressure and temperature over different parts of the earth.

In a series of papers on "World Weather", Sir Gilbert eventually succeeded in establishing three types of oscillations in the

general circulation of the atmosphere. By far the most important oscillation from our point of view was the Southern Oscillation, which has been described in an earlier chapter. The Southern Oscillation represents a tendency for high pressure over the Pacific Ocean to be associated with low pressure in the Indian Ocean and vice versa.

The discovery of the Southern Oscillation led Sir Gilbert to believe that a further search may reveal factors which have a close association with monsoon rainfall. In his early years as the Director General of the Indian Meteorological Service, Sir Gilbert outlined a scheme using eight factors for monsoon prediction. These factors are briefly described here for their historical interest.

(*i*) *Late and heavy snowfall over northern and western India during pre-monsoon months*

This was considered to be unfavourable for the monsoon. An unusually large snow cover in late spring generates a region of local high pressure over northwest India. Often this generates northerly winds which are unfavourable for monsoon rains over the Punjab and adjoining regions.

(*ii*) *Heavy rains at Zanzibar and Seychelles in April and May*

Sir Gilbert reasoned that heavy rains at Zanzibar and Seychelles was an indication of the predominance of an equatorial flow pattern over the monsoon. Consequently, this factor was also unfavourable for monsoon rain.

(*iii*) *High Pressure over South America*

South American pressure may be taken to represent conditions over the south Pacific Ocean. According to the Southern Oscillation, this should favour the formation of low pressure over the Indian Ocean and a strong monsoon current. This factor was, therefore, favourable for the monsoon.

(*iv*) *High pressure around Mauritius and Australia in spring*

These conditions again reflect a predominance of equatorial flow. Consequently, they were not favourable for a good monsoon.

(v) *High pressure over India during the previous year*

This factor was found to be favourable for a good monsoon, although the physical reason was somewhat obscure.

(vi) *Winds observed by ships in the Indian Ocean*

These winds, especially in the month of May, provided a good indication of an advance of the monsoon over India.

(vii) *The onset of summer rains over Ethiopia in May*

It was found that a large departure from the normal date of onset of summer rains over Ethiopia was also reflected in the date of onset of the monsoon over India.

(viii) *Height of the river Nile*

The height of the river Nile in Egypt often provided useful indications of the onset of summer rains over Ethiopia. The latter, in turn, provides an indication of the onset of the monsoon over India.

New factors for Seasonal Prediction

As the years went by, it was observed that the association between some of the factors selected by Sir Gilbert Walker and the Indian Monsoon was not so strong as initially believed. For example, high pressure over India during the previous year was at first considered favourable, but later the correlation coefficient between this factor and monsoon rainfall fell down to very low values. Eventually, after a series of further experiments, Sir Gilbert Walker in 1924 selected six factors for predicting monsoon rain in northwest India and the Indian peninsula. They are summarised in Table 12.2.

With the passage of years, it was observed that the performance of even these factors, selected in 1924 after considerable experimentation, was very variable as far as their predictive value was concerned.

One of the main predictors in Walker's technique was South American pressure. Until 1950 this factor showed a positive association with monsoon rainfall, but there were wide variations in the magnitude of the correlation coefficient during different

Table 12.2
Walker's Predictors for Summer
Monsoon Rain

	Factors	*Period*
I. Indian Pensinsula	1. S. American pressure	April—May
	2. South Rhodesia rain	October—April
	3. Dutch Harbour temperature	December—April
	4. Java rain	October—February
	5. Zanzibar rain	May
	6. Cape town pressure	September—November
II. Northwest India	1. S. American pressure	April—May
	2. South Rhodesia rain	October—April
	3. Dutch Harbour temperature	March—April
	4. Equatorial pressure	January—May
	5. Snow accumulation	May
	6. Cape Town pressure	September—November

decades. It varied from about 0.02 to 0.78 for the Indian peninsula and from 0.13 to 0.63 for northwest India. In the last decade the tendency also appeared to have changed sign.

In a similar manner, the association between Java rain and the Indian monsoon has shown wide variations in magnitude, and also a change in sign. The Dutch harbour temperature was omitted after a few years, because the station concerned was closed down. The equatorial pressure has also shown changes of sign twice, and the magnitude of the correlation coefficient is low.

In more recent times, Indian meteorologists have tried to improve Walker's prediction technique by incorporating new factors, whose association with the Indian monsoon appeared to be better than the ones originally selected by Sir Gilbert Walker. The range of temperature over Punjab and the Indo-Gangetic river discharges are some of the new factors that are included in the forecast formula for northwest India.

Another innovation has been the inclusion of upper air data in the prediction scheme. The upper winds of Bangalore, Calcutta and Agra have been included as new factors since 1956. In table 12.3 we present correlation coefficients between upper winds and monsoon rain for the three decades between 1921 and 1950.

Table 12.3
Correlation Coefficients between Upper winds of India and Monsoon Rain
(Correlation Coefficients are expressed in hundredths)

	1921—30	*1931—40*	*1941—50*	*Entire Period*
Peninsula				
1. Bangalore Northerly (6 km in April)	40	47	56	47
2. Calcutta Easterly (4 km in May)	—39	—19	—71	—43
Northwest India				
1. Agra Easterly (2 km in March)	02	37	66	35
2. Calcutta Easterly (2 km in May)	—62	—49	—71	—60

Table 12.3 illustrates the main difficulty of this technique. As we can see, values of the correlation coefficient show wide fluctuations over the decades. And, as mentioned earlier, there were a number of occasions when the correlation coefficient changed sign. The performance of the prediction formula has, therefore, suffered, because there is no known method to ascertain when and how the influence of a given predictor will change with time. It is also of interest to note that the overall prediction formula contains, eventually, a certain amount of uncertainty regarding its success. In statistical terms, it is possible to evaluate how often we may expect the prediction formula to turn out to be correct on the basis of mere chance. This turns out to be, approximately, about one in four. Over a long series of years, we may expect reasonable success on every four occasions out of every five.

Currently regression equations are used by the Meteorological Department of India to predict (a) the onset date for the summer monsoon over the southern tip of India and (b) the total rainfall from the beginning of June to mid-September. The latest predictors for the date of onset are :

—direction of mean January wind at 300 mb over Delhi

—direction of mean January wind at 200 mb
over Darwin in north Australia
—mean February wind direction at 200 mb
over Trivandrum and Madras
—the mean meridional component of wind at 200 mb
over Calcutta in December of the previous year.

These predictors were selected after applying statistical tests to screen a large number of other predictors. While a fair amount of success has been achieved in predicting the date of onset for a normal monsoon, the above predictors were not very successful in years when the monsoon was not typical. The variance of the date of onset over Kerala is approximately 8 days. When the onset is delayed by more than 8 days, as was the case in 1979, a regression equation was not very successful; otherwise, the root mean square error in prediction is about 2 days.

A different set of predictors is used for northwest India and the Indian peninsula. The rainfall in these two sectors is homogeneous in terms of high variability. The predictors currently in use are;

(a) Northwest India

—south American pressure (Buenos Aires) in April
—mean position of a ridge at 500 mb at approximately 12°N
and along 75°E in April
—mean equatorial pressure from January to May at Seychelles,
Jakarata and Darwin
—mean April temperature at Ludhiana.

(b) Indian peninsula

—south American pressures measured by departures from
normal at Buenos Aires, Cordoba and Santiago in April and
May
—the position of the axis of an upper air ridge along approximately, 12°N, 75°E at 500 mb in April.
—mean minimum temperature for March in Jaisalmer, Jaipur
and Calcutta.

Regression equations using the above predictors provide some indication of the total rainfall over these regions for a hundred day period from the beginning of June, but they give figures of

anticipated rainfall over a very broad region. They are not useful for indicating the likely rainfall distribution on a monthly basis. To overcome this difficulty auto-regressive models have been recently introduced by the Meteorological Department of India.

Autoregressive Integrated Moving Averages (ARIMA)

The heading, ARIMA, sounds a little complicated, but a description of this technique is appropriate, because monsoon meteorologists are now trying out this method for prediction of rainfall.

If we plot the rainfall values for every year, we will have what is known as a time series of monsoon rain. A time series represents a collection of observations, in sequence, at equal intervals of time. Many time series arise in economics and engineering. For example, we refer to the prices of shares for sucessive days or the total quantum of exports in successive months. Some instruments take measurements continuously and produce a continuous trace instead of observations at discrete intervals of time. In some laboratories, it is important to keep the temperature and humidity as constant as possible and instruments are installed to measure these variables continuously. Such records are examples of a continuous time series instead of a discrete one. When we analyse a time series, the first step is to plot the data and to see what the series looks like. In many economic time series we often encounter a regular seasonal effect, which exhibits a "high" in winter or a "low" in summer. Such a time series is dominated by seasonal effects.

In addition, there are oscillations or cycles that are revealed by a close examination of data. If we examine the temperature data of a given station, it would be readily seen that there is a fairly well defined diurnal cycle. The temperature would attain a maxima little after mid-day and a minima shortly after midnight. Economic data may be similarly affected by the presence of cycles with periods of five or seven years. If a cycle is found to dominate, then prediction becomes simple because we have only to identify what the cycle looks like at any given point of time.

Apart from cycles and periodicities, statisticians are concerned with "trends". A rough and ready definition of a trend may be a long term change in the mean value of the time series. The

difficulty with this definition lies in deciding what is mant by "long term". If, for example, we were interested in climatic changes over a very long period, such as 100 years, but if one had only 20 years' data, the long term oscillation of 100 years would appear to be only a trend. On other hand, if we had several hundred years of data, the long term oscillation would become more visible. An assessment of what is "long term" is necessary.

If we could remove the trend and seasonal variations from a set of rainfall data, we will be left with a sequence of residuals. The main problem in long range prediction is to identify what physical processes determine the residuals. If a set of physical laws could be found to govern the residual, then one might, with confidence, state that the time series was determinate. By this we mean that if we could start from a given point in time, we could predict what the future would look like by applying the physical laws of atmospheric motion. If the physical laws cannot be identified, then one might conceivably argue that the future behaviour of the residuals would be determined by their past history. In other words, the behaviour of the series would be stochastic in nature.

The question of determinism against the stochastic nature of atmospheric motion has not been finally resolved. Modern developments in computer technology and numerical modelling suggest that a good part of atmospheric motion is determinate. If we apply the principle that atmospheric systems which generate weather, such as, tropical cyclones or other types of vortices on different scales of motion, conserve their spin or, vorticity as the meteorologists define spin, then the future location of these systems could be determined with the help of computer-oriented models. But, if we consider motion on a longer period of time, such as the interannual variability of monsoon rain, it is by no means clear that the process is determinate. The stochastic nature of climatic variations needs consideration when problems of this nature are encountered.

A time series is stationary if there is no change with time in its broad statistical features. In other words, if we compute the mean rainfall of a given station or a given region, with, say, twenty years data, then will the mean remain unchanged if we enlarge the time series to contain fifty or hundred years of data ?

Unfortunately, this is not the case with most atmospheric variables. We cannot assert that a monsoon time series depicting rainfall is stationary.

Despite this difficulty it is possible, within reasonable limits of confidence, to assume that a part of the time series is "auto regressive". This means that a good portion of the time series is determined by its past history.

To what extent this is true may be estimated by determining the correlation that exists, for example, between the rainfall in the previous year with rainfall of the year before. Similarly correlation coefficients could be worked out between the rainfall of a given year with those of the preceding two, three, four or five years. This statistical association is known as an auto-correlation coefficient. By computing auto-correlation coefficients, we measure the impact of past history on the behaviour of the time series. Apart from an autoregressive process, it is possible to eliminate, to some extent, the impact of trends by using a device which statisticians refer to as "moving averages". This implies taking the average of small segments of the time series. What we are trying to do here is to consider not the rainfall recorded each year but to plot, say, average of rainfall for five year periods. By this means we eliminate any trend that might exist in the time series over segments of five year duration.

We have described how to determine the past behaviour of a time-series. This is important for anticipating the future, but what is equally important is to find out what the residuals are going to do in future. Here the technique is to assume that the residuals represent a series of "shocks", which are random, on the time series after the trends, cycles and seasonal features have been taken care of. The distribution of random shocks which best fits the behaviour of the time series may be determined by a device that is described as a "best fit by least squares". The principle here is to try and generate a mathematical formula which will fit the behaviour of these random shocks. The formula which gives the best fit is computed by minimising the square of the difference between the shocks predicted by our formula and the actual observed values. One might wonder why we need to minimise the square of the difierences rather than the differences themselves. The reason for this procedure is to minimise the

magnitude of the difference rather than consider their signs. Some differences may be positive while others may be negative.

Such are the principles of the ARIMA technique. It has been tried for long range prediction of monsoon rainfall for the last two years in India. The results have been encouraging so far, and the average percentage of success appears to be around 65% for a forecast one month in advance. This is better than what might have been achieved by pure chance. An interesting feature of ARIMA lies in the fact that monsoon rainfall appears to be significantly influenced by the positions of a ridge at, approximately, 12°N and 75°E at 500 mb in April.

World Climate Programme (WCP)

Recently the World Meteorological Organization (WMO) and International Council of Scientific Unions (ICSU) have realised the need for studies on climate and climatic variability. This emphasis assumes considerable importance for India in view of our dependence on monsoon rains. It is to be realised that rainfall for us is a natural resource and, although it does not confirm to our usual idea of resources, the fact is that rainfall affects not only our agriculture but also the levels of our reservoirs that provide electricity and power. It affects the flow in our rivers and a wide variety of activities that have a large impact on our daily lives. A change in the pattern of monsoon rainfall could lead to a major redistribution of our wealth.

A World Climate Conference was convened by the World Meteorological Organization in February 1979 in Geneva. The Conference was attended by nearly 400 experts. This Conference was instrumental in drawing up two major objectives of the World Climate Programme :

—to improve our knowledge of the natural variability of climate and the effects of climatic change due either to natural causes or to human activities ; and

—to assist decision-makers in planning and coordinating climate-sensitive activities of economic, environmental and social significance so that these are less vulnerable to climatic change and variations.

The World Climate Programme, as it stands today, has four components : a Climate Data Programme, a Climate Applica-

tions Programme, a Climate Impact Study Programme and a Climate Research Programme. Each of these four components is important for the study of the monsoon. Of great importance is a study of possible changes that might be induced as a result of human activity. It has been conjectured that large scale climatic changes might be induced by an increase in the amount of Carbon dioxide into the atmosphere. The reasoning is that Carbon dioxide has little impact on solar radiation that enters the atmosphere, but it absorbs strongly the long wave radiation emitted by the earth in a narrow wavelength band. Consequently, it acts as a "greenhouse" by trapping longwave outgoing radiation. Current techniques suggest that substantial global warming and changes in climate could arise from the expected increase in Carbon dioxide content. The impact of Carbon dioxide increase would be largest near the polar regions because of the higher reflective properties of snow and ice. This could then bring about a change in the equator to pole temperature contrast which, in turn, could affect global monsoons.

While it is generally agreed that Carbon dioxide in the atmosphere has increased by 15% during the last century and is at present increasing by about 0·4% per year, it is possible that some effects of this may become detectable before the end of this century, and may become even significant by the middle of the next century.

The increasing use of energy and release of heat could also cause local climatic changes, especially in densely populated and heavily industrialised regions. Other human activities, such as, deforestation and increased used of nitrogen fertilisers could also have climatic consequences. These effects require further research and study. Degradation of our environment is a major concern for society because it could influence climate in other parts of the world. There is need today for cooperation between nations to avoid misuse of water resources, forests and for the preservation of fertility of soils. The world community of meteorologists is becoming conscious of the importance of climatic variations, especially as they relate to the monsoons of tropical lands, because the latter carry the biggest burden of the world's population.

CHAPTER XIII

FLOODS, DROUGHTS AND FAMINES

WE DO NOT wish to begin this chapter on a depressing note, but the damage and devastation that is caused every monsoon season by natural calamities makes it appropriate to discuss these events. Let us begin with floods and ask ourselves : why do we have floods ?

Floods

The flood plain is a low land that borders a river. It usually consists of alluvium and sediments that are carried and deposited on the river bed. Alluvium is rich soil which is beneficial for crop production ; consequently, in some river valleys floods have been turned to economic advantage. An example is Egypt, where the river Nile has provided rich alluvium. But, while Egypt has benefitted from the waters of the Nile, floods have been the scourge of India, China and several countries of the far east.

In our country, major floods are associated with the Ganga and the Brahmaputra rivers. Of the other rivers that are vulnerable to floods are the rivers of Orissa and Uttar Pradesh, along with the Narmada and the Yamuna. Heavy sedimentation is a feature of the Ganga and the Brahmaputra. It is believed that these large deposits of sediments have their origin in the Himalayan mountains. Owing to massive assaults on forests, the top soil is washed away by rainfall into these two major river systems, namely, the Ganga and the Brahmaputra. Persistent deposition of sediments on the river bed has reduced the water holding capacity of these mighty rivers, so that they are no longer able to withstand an additional water load generated by spells of heavy rain.

It is possible to mitigate, to some extent, the damage due to floods by constructing canals, flood control structures and dams. This is largely an engineering problem but a word of caution would be in order. Many of the worst disasters in our country have been associated with the burst of dams following heavy rain. The catastrophic events after the failure of the Machhu dam in Gujarat during August 1979 is a good example. To guard against such disasters, an efficient flood warning system is sine qua non for disaster mitigation. Along with this, we need research on estimating the maximum possible precipitation at strategic points of a river catchment.

Meteorological systems

Extreme rainfall events may be considered under three categories :

—Cloud bursts
—Monsoon depressions
—Tropical cyclones

A cloud burst represents highly concentrated rainfall over a small area lasting a few hours. They lead to flash floods, that is a flood which occurs suddenly, without warning, and lasts for a short time. There have been several examples of flash floods in our country. A recent example was the flood caused in the Luni river of Rajasthan in July 1979. A number of stations in Rajasthan recorded more than 10 cm of rain on July 16. A record rainfall of 29·4 cm was recorded at Ajmer on July 16. Apart from extensive inundation, a railway bridge over the Luni river was washed away. Oddly enough, while Gujarat and Rajasthan were devastated by floods, the remaining parts of the country were under the grip of a severe drought on account of a comparatively poor monsoon in 1979.

The exact mechanism which generates a cloud burst is imperfectly understood. Research suggests that they are manifestations of intense vortices on a small scale. These vortices generate strong convective currents which lift the moisture laden air with sufficient rapidity to enable them to shed their water load with great strength and ferocity. There is no satisfactory technique for anticipating the occurrence of a cloud burst because

of their small scale. One would need a very fine network of radars to be able to detect the likelihood of a cloud burst, and this would be prohibitively expensive. Thus we have to contend with the fact that disaster mitigation measures against cloud bursts and flash floods are going to be extremely difficult. But much can be done by way of identifying areas and meteorological situations that favour the occurrence of a cloud burst. In India, during the monsoon, such situations are created by the westward passage of monsoon depressions and mid-tropospheric low pressure systems.

We have described monsoon depressions in earlier chapters. When a depression forms in the Bay of Bengal, the pressure begins to fall over a wide region along the eastern coast of India. The wind, in response to the pressure gradient, picks up a cyclonic circulation which, in the northern hemisphere, is anticlockwise. Viewed from satellites, these depressions appear to be small vortices embedded in an atmosphere spinning round the earth.

On an average, about eight cyclonic depressions move from the Bay of Bengal into the land area between June and September. As the depression moves westwards from the head of the Bay of Bengal, a belt of rainfall extends to the southern and southeastern parts of Bengal and lower Assam. With the further movement of the storm westward, the rain belt extends to Orissa, Chota Nagpur and Bihar. By the time the storm crosses the Orissa coast and enters Madhya Pradesh, the presence of the depression strengthens the Arabian Sea branch of the monsoon. This causes another spell of moderate to heavy rain over Madhya Pradesh and the southern parts of Uttar Pradesh, as well as over the north Peninsula. The rainfall may then be carried by the depression into Rajasthan and Gujarat before it merges with the seasonal low pressure area over northwest India.

Sometimes these depressions recurve towards the north and eventually break up over the Sub- Himalayan regions of Punjab and Kashmir. In such a situation, the Arabian Sea branch of the monsoon feeds extra moisture into the the storm and very heavy rain is recorded over the hilly districts of Punjab and Himachal Pradesh.

It has been observed that after the passage of a depression, the monsoon weakens and the rains slacken. However, after a gap

of few days, the monsoon tends to revive and another depression forms at the head of the Bay of Bengal. The pattern of rainfall associated with the preceding depression is then repeated.

Cyclonic Storms of the Monsoon

Sometimes the fall of atmospheric pressure is much larger than what is observed in a depression. The region of low pressure is more localised and the air acquires much greater cyclonic spin. A depression is then said to concentrate into a tropical storm or cyclone. Tropical storms are not frequent during the monsoon season.

The intensity of a tropical storm may be conveniently expressed in terms of the velocity of wind. The procedure is based on an early method of indicating the velocity of wind devised by Admiral Beaufort in 1905. He divided wind velocities into several categories according to their effect on objects at sea. In agreement with Beaufort's classification of wind velocities, depressions and tropical cyclones are described in Table 3.2 of Chapter III.

Table 13.1
Frequency of Cyclonic Storms in the Bay of Bengal and Arabian Sea—1891-1960

Month	*Bay of Bengal*	*Arabian Sea*
January	4(1)	2(0)
February	1(1)	0(0)
March	4(2)	0(0)
April	18(7)	5(4)
May	28(18)	13(11)
June	34(4)	13(8)
July	38(7)	3(0)
August	25(1)	1(0)
September	27(8)	4(1)
October	53(19)	17(7)
November	56(23)	21(16)
December	26(9)	3(1)
Total	314(100)	28(48)

The figures within brackets indicate the number of Storms which reached "severe" intensity.

A considerable amount of literature is now available on the frequency of tropical storms during the monsoon months. In the following table we show the number of storms that formed in the Bay of Bengal and the Arabian Sea in different months during the seventy year period from 1891 to 1960.

Interesting features are brought out by this table. Looking at the distribution figures, we find that the number of storms in the

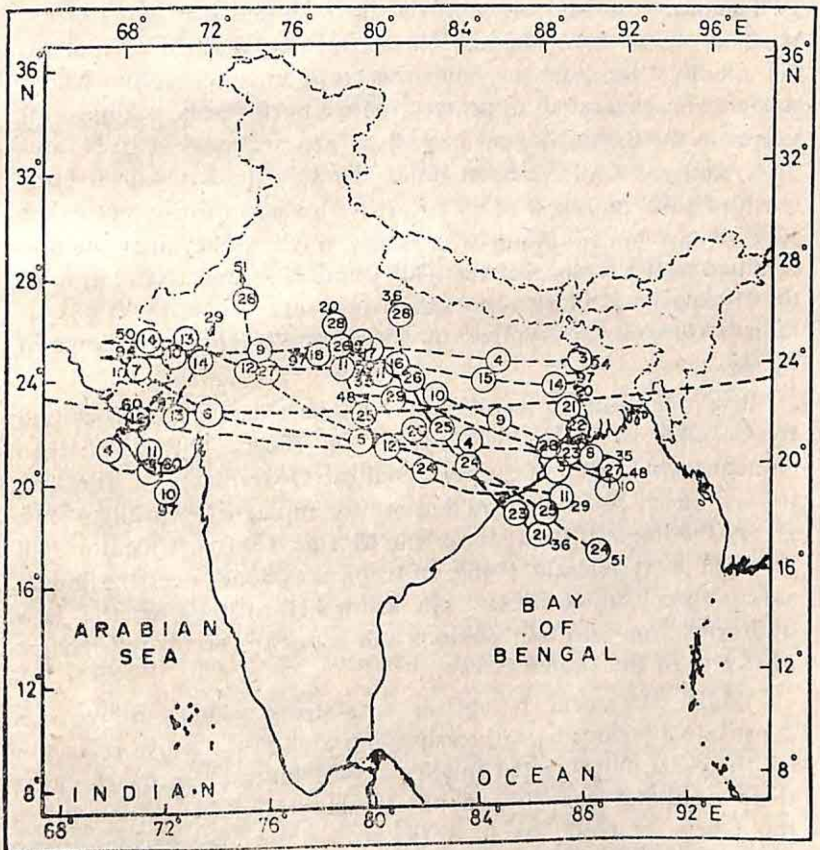

Fig. 13.1—Normal tracks of cyclonic storms in July. Figure inside the circle indicate day of the month. The years are given at the beginning and end of the tracks.
The territorial waters of India extend into the sea to a distance of twelve nautical miles measured from the appropriate base line. Responsibility for the correctness of internal details shown on the map rests with the publisher.

Bay of Bengal is much greater than over the Arabian Sea. The maximum number of storms in the Bay of Bengal occur in the months of October and November. In the Arabian Sea the largest number of storms is observed in May, June, October and November. The early part of the moosoon season is favourable for the formation of tropical storms in both the Bay of Bengal and the Arabian Sea.

The month of May shows a rise in the number of Bay storms. Most of them have their origin between 10°N and 15°N during the month. But, as the monsoon sets in, the region where storms are generated appears to move northwards. Almost all storms in the Bay of Bengal have their genesis between 16°N and 21°N and west of 92°E in June. By July the Bay storms form north of 18°N and west of 90°E. It is also noteworthy that most July stroms move along a westerly track. They are generally confined to the region between 20°N and 25°N and recurvature to the Himalayan foothills is comparatively rare. The tracks of cyclonic stroms for a typical monsoon month (July) are shown in fig. 13.1.

It is interesting to recall that very severe damage was caused to Calcutta by a cyclonic strom in 1864. The disasterous consequences of this storm prompted the Government of the day to set up a Meterological Serivice for India. Even prior to this, Henry Piddington, President of the Marine Courts, Calcutta had initiated a systematic study of tropical cyclones over the Indian seas. His celebrated book "The Sailor's Horn Book for the Law of Storms" ran into four editions and is one of the earliest records of storms in the Indian Ocean.

While the storm is out at sea, strong winds and high seas constitute a serious hazard to shipping. When the storm is nearer to the coast, unprotected smaller vessels and fishermen, who earn their livelihood from the sea, are in danger of being caught in the storm. Finally, as a storm crosses the coast it is known to cause tidal bores and surges which inundate coastal areas and cause considerable loss of human lives and property. Heavy rains associated with a storm have been known to cause floods of large dimensions. From newspaper reports it was estimated that about 700 lives were lost when a severe cyclone struck the coastal town of Masultipatnam in October 1949. The total damage to crops

extending over an area of 9·4 lakh acres was estimated to be about Rs. 4·6 lakhs.

Every meteorological service has a cyclone warning system designed to warn the public against an approaching storm. Warning signals were first introduced in India in 1865 at the port of Calcutta. Today this system has been extended to cover all the major ports on the eastearn and western coasts of India. The major storm warning centres of the country are located at Calcutta, Madras and Bombay. The location of a storm, its probable intensity and future movement is analysed at these centres, and a warning is sent immediately to all user interests who need to be forewarned. The smaller ports along the coast are instructed to hoist signals which warn nearby ships of an approaching storm. The nature of the signals is altered from time to time to indicate the intensity of the storm, its proximity to a port and the expected deterioration in weather. The general public of the coastal regions are also advised from time to time by radio and other means about the danger of tidal waves, gales, heavy rain and other hazards associated with a tropical cyclone.

There has been much discussion on the type of protective measures that could be profitably taken up in regions that lie on the normal path of cyclones. We have in mind measures of a permanent nature, such as, the construction of barriers and high embankments. Such measures could also include the construction of surge-proof buildings that may withstand high wind velocities, and elevated platforms for shelter against floods. It has been reported that in some parts of the world, such as the Netherlands, the local population are asked to seek refuge on earthen mounds in the event of a flood.

Preliminary studies reveal that the construction of a permanent embankment requires considerable engineering skill and a large financial investment. A rigid construction, for example, made up of concrete may not stand up to the pounding and suction of large waves for a considerable length of time. We have to think in terms of flexible constructions with a protective layer designed to prevent the main structure from being washed away.

The Structure of Cyclonic Storms

One of the most interesting features of the tropical cyclone, apart

from its strong revolving motion and heavy precipitation, is its 'eye'. The 'eye' refers to the central region of the cyclone. It is marked by comparatively calm conditions and, sometimes, by clear skies. In India, an interesting observation was made concerning the 'eye' of the Masulipatnam cyclone of October 1949. The following is an extract of the observations recorded by an observer at Eluru, about 40 miles northwest of Masulipatnam during the period of the strom :

"The worst phase of the Storm was experienced between midnight and 3 A.M. on the 28th. The wind speed during this period might have been easily of the order of 130 to 145 kilometres per hour (80 to 90 mph). The wind speed subsided after 4 A.M. and by 6 A.M. had considerably abated and was only a few miles per hour. By 7 A.M. the wind strengthened again, the direction backing to westsouthwest (WSW). By 8 A.M., it was again blowing gales and this continued with unabated fury till 2 P.M. The maximum wind speed during this period might have been of the order of 80 to 100 kilometres per hour (50 to 60 mph). The winds subsided by 4 P.M."

The period of calm winds between 4 A.M. and 7 A.M. on the 28th clearly represented the passage of the "eye". With the help of other data, it was estimated that the diameter of the calm region was of the order of 15 km.

Theoretical reasoning leads us to infer that air which is spinning round the centre of a cyclone should conserve its angular momentum. If r is the radial distance from the storm centre and v is the tangential velocity of winds, this means that the product of r and v ($r \times v$) should be constant. But, this would mean that as the air approached the centre, with decreasing r, the velocity of the wind (v) would increase to an infinite magnitude. This cannot happen because the kinetic energy of the storm has an upper limit. Consequently, being unable to force its way to the centre of the storm, the spinning air begins to rise upwards instead of rotating round the storm centre. This is how an 'eye' is formed.

Aerological soundings with the help of radiosondes and research aircraft have revealed that the air above the central region of the cyclone is much warmer than the sounding air. An idealized vertical cross-section showing the temperature distribu-

tion in a tropical storm is reproduced in fig. 13.2. The figure represents an overall picture of the structure of a tropical storm based on the study of a large number of cyclonic storms in the north Atlantic and the Pacific.

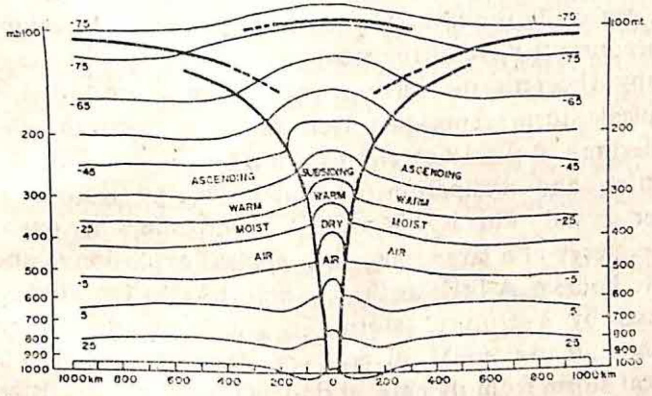

Fig. 13.2—Idealized vertical cross-section showing the temperature distribution (°C) through a tropical cyclone. Note that the atmosphere in the vicinity of the centre of the storm is warm and dry.

Besides the warm air in the 'eye' of the storm, it is interesting to note that the tropical cyclone occupies a wide region of several hundred kilometres. The air within the 'eye' is subsiding, warm and dry. On other hand, the air outside the 'eye' is a column of ascending warm moist air. The walls of the 'eye' represent zones of considerable shearing motion.

The wind field in the mature tropical storm indicates strong spiralling inflow of air with counter clockwise rotation at lower levels. At higher levels of the troposphere an outflow of air is observed with clockwise rotation. In the central region, however, the rotation is counter-clockwise even at upper levels. These observations suggest that more air is removed by an outflow at upper levels than is brought in at lower levels. The fall in pressure that is observed at the surface represents an excess of upper outflow over low level inflow. The central problem of cyclone formation is to explain how the upper outflow is maintained throughout the lifetime of a cyclone.

Several attempts have been made to find a correlation between upper tropospheric flow patterns and storm development. There seems to be some amount of agreement that the existence of an

upper anticyclone favours the formation of a cyclone. But at best this is only one of the necessary conditions for cyclone formation, because there have been situations when upper anticyclones have failed to generate a storm. There is need today for more precise information on the early stages of the development of an incipient storm, especially the formation of a warm central region which later becomes the eye of the storm.

Many theoreticians have drawn our attention to the fact that a tropical storm represents tremendous generation of energy. Considering a storm to extend over a radius of 660 km, the total production and dissipation of kinetic energy is about 15×10^{18} ergs per second. This is $0{\cdot}36 \times 10^{12}$ kilowatt-hours per day.

The energy of a large (megaton) nuclear explosion is about 10^9 kilowatt hours. A little arithmetic tells us that the kinetic energy generated by a tropical storm is worth about 360 megaton explosions. If we could divert even a small part of the energy of a tropical storm from its path of destruction, there is little doubt that the benefits will be substantial.

Indeed this appears to have been the motive behind a series of experiments which were performed in the United States. Scientists argued that if the wall of clouds round the eye of a mature storm were seeded with Silver Iodide, it would release considerable amounts of latent heat. This could well upset the delicate balance of forces within the storm. The end result would be a decrease in the maximum wind speed and, as a consequence, a reduction in the destructive power of the storm.

This hypothesis was tested on Hurricane Esther in September 1961 and Hurricane Beulah in August 1963. It was reported that in both experiments the results appeared to support the hypothesis. In one of the storms, the average reduction in wind speed was about 14 per cent. Admittedly, one can hardly draw firm conclusions on the basis of two experiments, but considering the wild destructive power of these systems, it would be worthwhile to try and see if we could tame nature a bit, even if it be by a minute fraction. Certainly the resources of nations have been spent, or wasted, on much less profitable ventures in the past.

Rainfall Analysis

Tropical cyclones are known to generate heavy rain ; the one

which struck the coast of Orissa between October 26 and 31 in 1971 produced rainfall amounts varying from 24 to 30 cm a day over large areas of Orissa and West Bengal. There are similar examples of heavy rain associated with tropical cyclones in Bangladesh, Burma, Sri Lanka and other countries in southeast Asia. Cyclone Tracy produced 82 cm of rain over Darwin in northern Australia 1974. Another cyclone, Althea, which hit the coast of Queensland in Australia in December 1971 recorded rainfall upto 25 cm in a 24-hour period.

In table 13.2 we reproduce some of the world's record rainfalls from a publication by Dr. Linsley and his collaborators in 1975.

Table 13.2

World's greatest observed point rainfalls

(Linsley et al, 1975)

Duration	*Depth in*	*mm*	*Location*	*Date*		
1 min.	1·50	38	Barot, Guadeloupe	Nov.	26	1970
8 min.	4·96	126	Fussen Bavaria	May	25	1926
15 min.	7·80	198	Plumb Point, Jamaica	May	12	1916
20 min.	8·10	206	Curtea de Arges, Rumania	July	7	1889
42 min.	12·00	305	Holt, MO	June	22	1947
2 hr. 10 min.	19·00	483	Rockport. W. Va	July	18	1889
2 hr. 45 min.	22·00	559	D'Hanis. Tex	May	31	1935
4 hr. 30 min.	30·80	782	Smethport, Pa	July	18	1942
9 hr. 00 min.	42·79	1087	Belouve, Reunion	Feb.	28	1964
12 hr. 00 min.	52·76	1340	Belouve, Reunion	Feb.	28-29	1964
18 hr. 30 min.	66·49	1689	Belouve, Reunion	Feb.	28-29	1964
24 hr.	73·62	1870	Cilaos, Reunion	March	15-16	1952
2 days	98·42	2500	Cliaos, Reunion	March	15-17	1952
15 days	188·88	4798	Cherrapunji. India	June July	24— 8	1931
31 days	366·14	9300	Cherrapunji, India	July		1861

Not all the rainfall in the above table was associated with tropical cyclones, but it is indicative of the tremendous amounts of rain that have been generated all over the world. It is to be noticed how very large amounts of rain are generated within short intervals of time. The table shows, for example, rainfall

amounts of the order of 13.4 cm on the island of Reunion in 12 hours. This is roughly the amount of rain which western parts of Rajasthan receive in a whole year.

Most meteorological services have been collecting rainfall data extending over a large number of years. Hydorologists like to express the frequency of rainfall in terms of 'return periods'. A return period is the reciprocal of the frequency. Thus, a flood with a fifty-year period would signify that a flood of a given magnitude was likely once in fifty years. The length of record at individual stations usually ranges from 5—50 years. In India, the longest rainfall records exceed seventy years.

There are statistical techniques by which one could compute the likelihood and the corresponding magnitude of rainfall amounts for different parts of the country. These techniques are beyond the scope of this book and will not be discussed, but their importance lies in flood forecasting which is a problem in hydrology. For the present, we may note that flood forecasting in India is now beginning to make use of model catchments. These models strive to locate the sources of discharge of water by way of flow from the river and its tributaries. The model then seeks to convert the input of water in the form of rainfall into a flood discharge. This is achieved by a synthesis of past floods from the individual rainfall that cause them. More recently, attempts are in progress to develop mathematical models that convert rainfall directly into flood discharges on a computer.

Droughts and Famines

We have so for discussed a precipitation extreme. This was concerned with heavy rainfall leading to an excess of water and consequent floods. But during the monsoon we experience another precipitation extreme, which is a 'drought' or a distress situation caused by lack of rainfall. The failure of rains may be viewed from two aspects : firstly, the rainfall may be insufficient or, secondly, it may be sufficient for the season as a whole but with a wide gap separating two or more spells of rain. The quantum as well as the timeliness of rainfall is important.

Scientists have tried to define a distress situation caused by lack of water into three categories of drought depending on the meteorological, hydrological and agricultural aspects. In general terms, they are ;

(a) *Meteorological* : Meteorological drought is a situation when the actual rainfall is significantly less than the climatologically expected rainfall over a wide area.
(b) *Hydrological* : Hydrological drought is associated with marked depletion of surface water and consequent drying up of lakes, rivers and reservoirs. A hydrological drought follows, if a meteorological drought is sufficiently prolonged.
(c) *Agricultural* : Agricultural drought occurs when the soil moisture is inadequate to support a healthy growth of crops to the stage of maturity.

There is not much by way of meteorological observations before 1875, but a broad account of famines in India is available. It appears that from the beginning of the eleventh century up to the end of the seventeenth century, there were as many as fourteen famines. Not much is known about these famines but from all accounts, it appears that the resilience of the country as a whole did not falter whenever distress was caused by a famine. Thus, there are examples to indicate that whenever a part of the country was affected by a famine, there was a serious attempt to assist the affected parts by diverting food grains from surplus regions to areas affected by a drought. But, unfortunately, poor communications often came in the way of timely relief. Thus it happened that foodgrains that were sent to the affected regions on the east coast of India often reached the region after the next year's monsoon had already set in. Lack of storage facilities sometimes led to wastage that could have been avoided. In table 13.3 we produce some of the major droughts that have affected the country in the last hundred years.

Table 13.3

Major famines in the last hundred years

(After Sarker, 1979)

Year	*Area affected*	*Percentage of total area of the country*
1877	Sub-Himalyan West Bengal, Bihar plains, East Uttar Pradesh, West Uttar Pradesh, Haryana, Himachal Pradesh, Rajasthan, Madhya Pradesh, Gujarat, Konkan; Marathwada, Vidarbha; Coastal Andhra Pradesh, Telangana, Jammu & Kashmir.	63.9

1883	West Uttar Pradesh, Haryana, Himachal Pradesh, Rajasthan.	20.0
1891	North Assam, Sub-Himalayan West Bengal, Bihar plains, Rajasthan, Andhra Pradesh, Tamilnadu.	33.6
1899	Haryana, Punjab, Himachal Pradesh, Madhya Pradesh, Rajasthan, Gujarat, Saurashtra, & Kutch, Maharashtra, Andhra Pradesh, Tamilnadu, Coastal Karnataka, Kerala.	64.5
1901	Sub-Himalayan West Bengal, Orissa, Punjab, Rajasthan, Gujarat, Vidarbha, Saurashtra & Kutch, South Interior Karnataka.	29.6
1904	Sub-Himalayan West Bengal, East Uttar Pradesh, Punjab, Gujarat, Vidarbha, Andhra Pradesh, West Rajasthan, Saurashtra & Kutch and South Interior Karnataka.	52.6
1905	West Uttar Pradesh, Haryana, Punjab, East Rajasthan, Konkan, Madhya Maharashtra, Marathwada, North and South Interior Karnataka.	29.8
1907	East Uttar Pradesh, West Uttar Pradesh. Haryana, Punjab,. Himachal Pradesh, East Rajasthan, West Madhya Pradesh.	25.2
1911	Punjab, Rajasthan, Gujarat, Madhya Maharashtra, Telangana, Saurashtra & Kutch.	24.7
1918	East Uttar Pradesh, West Uttar Pradesh, Haryana, Punjab, Himachal Pradesh Rajasthan, West Madhya Pradesh; Saurashra & Kutch, Gujarat, Maharashtra, Telangana, Rayalaseema, Jammu & Kashmir.	70.7
1920	Punjab, West Madhya Pradesh, Gujarat, Konkan, Marathwada, Vidarbha, Coastal Andhra Pradesh, Gujarat, Telangana, West Rajasthan, Jammu & Kashmir.	36.4
1939	Haryana, Punjab, Rajasthan, Gujarat, Telangana, Saurashtra & Kutch.	22.9
1941	West Uttar Pradesh,Haryana,East Madhya Pradesh, Konkan, Telangana, West Rajasthan.	23.1
1951	Bihar Plains, Haryana, Punjab, Rajasthan, West Madhya Pradesh, Gujarat, Kerala, Saurashtra & Kutch, Jammu & Kashmir.	37.2
1965	East Uttar Pradesh, Haryana, Punjab, Himachal Pradesh, East Rajasthan, Madhya Pradesh, Gujarat, Vidarbha, Kerala, Jammu & Kashmir.	40.8
1966	Gangetic West Bengal, Bihar, East Uttar Pradesh, East Rajasthan, Madhya Pradesh.	30.2

1971	South Assam, Madhya Maharashtra, Vidarbha, Telangana, West Rajasthan, South Interior Karnataka, Jammu & Kashmir.	29.6
1972	South Assam, Sub-Himalayan West Bengal, Bihar Plains, Punjab, East Rajasthan, Gujarat, Maharashtra, Andhra Pradesh, Kerala, Saurashtra & Kutch, South Interior Karnataka, Jammu & Kashmir.	47.6
1974	Orissa, Punjab, East Madhya Pradesh, Gujarat, Madhya Maharashtra, Marathawada, Vidarbha, West Rajasthan, Saurashtra & Kutch, South Interior Karnataka, Jammu & Kashmir.	42.8

From table 13.3 we see that the three worst famines were in 1877, 1889 and 1918 when over 60% of the country faced acute distress. In an interesting article on Indian crop failures, Dr. McAlpin suggests that there was a decline in rainfall between 1886-1897 and between 1898-1906 over western India. As we have noted earlier, there are short period variations in rainfall, but they are not indicative of any systematic trend. Consequently, it is difficult to devise a prediction technique merely on rainfall periodicities. It must be clearly realised that while a period of low harvest is largely the consequence of poor rainfall, a famine need not be so ; on most occasions, a famine could be avoided by administrative procedures by way of advance planning for food distribution. The relationship between the distribution of rainfall and crop yields is extremely complex. We have, however, to contend with the fact that rainfall is uncertain over many parts of India. The coefficient of rainfall variation puts sufficient stress on this aspect. It is possible to overcome, to some extent, the uncertainty of rainfall by adequate changes in cropping patterns. Thus, it may be possible to select weather resistant crops. M.P. McAlpin notes that in western India, weather resistant varieties of some crops were developed as early as 1840. A major cotton plant was found to be one that had a long tap-root which helped to retain moisture during a break in rains. Similar improvements in cropping patterns have been now introduced over many parts of our country.

It must be noted that there has been very considerable

progress in irrigation in the last three decades. The total irrigated area is shown in table 13.4. The total area under irrigation has increased by a factor of 2.5 in the last 30 years. There is certainly scope for more irrigation, but even so it is doubtful whether irrigation alone would make us completely independent of monsoon rains. Along with progress in irrigation, emphasis must of necessity be placed on better prediction techniques for monsoon rainfall.

Table 13.4

Progress of irrigation

(Gross irrigated area in million hectares)

	Major and medium	Minor	Total
Pre-Plan (1950-1951)	9.7	12.8	22.6
First Plan (1955-56)	11.0	14.1	25.1
Second Plan (1960-61)	13.1	14.8	27.9
Third Plan (1965-66)	15.2	17.0	32.2
Annual Plans (1968-69)	16.9	19.0	35.9
Fourth Plan (1973-74)	18.8	23.5	42.3
1977—78	24.8	27.3	52.1

There are three major causes of droughts as far as rainfall is concerned :

—Late onset and early withdrawal.

—Lean rainfall due to paucity of depressions and low pressure systems and

—Prolonged breaks in monsoon rainfall.

We do not fully understand the physical mechanisms which lead to the above meteorological situations. Improvements in prediction lean heavily on our monitoring facilities. The latter should improve when the Indian National Geostationary Satellites are in orbit. This will provide some indication of the onset of the monsoon. It is not unreasonable to hope that with more information on cloud structures, one may be able to anticipate conditions that are likely to lead to a late or an early onset.

The second aspect is one concerned with a global view of weather. The present information which we have as a conse-

quence of the Monsoon Experiment (MONEX) of 1979 suggests that there are meteorological telecommunications that link up the passage of depressions and low pressure systems with events occurring in other parts of the world. The Southern Oscillation, which was described in earlier chapters, is an example of a meteorological link on a global scale.

The last aspect, namely, prolonged breaks is again intimately linked with the dynamics of the monsoon. Past rainfall data suggest that prolonged breaks in monsoon rainfall have a tendency to occur towards the second half of the season, that is, in August and September. Thus, in 1974 and in 1979 most parts of northwest India suffered a break in monsoon rains for periods ranging from six to eight weeks. A prolonged break of nearly six weeks also occurred in 1981. The evidence that we have suggest that these breaks are sometimes linked with a quasi-stationary anticyclonic circulation that establishes itself over northwest India. This anticyclonic circulation inhibits the upward motion of air and there is less chance of rainfall as a consequence. What meteorological features lead to this type of circulation is not yet well understood.

There are problems of this nature on which meteorologists are now engaged. With more data coming in, and with greater awareness of the importance of prediction, it is expected that new results will emerge in the coming years which will improve our capability of anticipating these events.

CHAPTER XIV

THE ECONOMICS OF MONSOONS

INDIA IS ONE OF THE few countries of the world where the rainfall is seasonal ; consequently, its crop production is very sensitive to monsoon rains. It has been estimated that at the present rate of growth of population our annual requirement of food grains will be neary 200 million tonnes around 2000 A.D.. To achieve this figure, an annual growth rate of around 5% in food production is needed to maintain the present level of nutrition. But our aim should be to improve the present level, which means a higher annual growth rate is necessary. Unfortunately, one bad monsoon, as in 1979, could lower the crop output by as much as 10-15%. As we have pointed out earlier, the regional distribution of rainfall over the country is uneven and shows wide variations. The map depicting the coefficient of variation clearly brings out this feature. The regions which are vulnerable to famines and droughts are located where the rainfall variation is high. These are also the regions where the annual rainfall amount is between 35 and 100 cm. The failure of rains is less common in regions which have a small variability of rainfall ; they are also the regions of high rainfall.

In a country whose economy is developing rapidly, the demand for food is linked with many economic considerations. The rise in the average income of citizens is, for example, a reason for an increasing demand for foodgrains. Notwithstanding a distinct improvement in living conditions, a large portion of the population needs better nourishment. At present it has been estimated that the annual increase in agricultural production is around 3% which is a little below the estimated demand of 5%. The figures

suggest that greater agricultural productivity is necessary for improving the nourishment provided to each citizen.

Agricultural engineers suggest that there is considerable scope for increasing the productivity of our land. Thus, in 1962-63, it was estimated that the average production of rice in India was 1·80 kg per hectare against a figure of 5·26 kg in Japan. Similar figures are available for wheat and other crops. Increase in productivity could be brought about by a greater input of fertilisers. In a book on "Famines in India", B.M. Bhatia suggests that even a modest increase in the application of fertilisers could bring about an increase in foodgrain production between 35 and 55%. In recent years an increasing trend is observed in the use of fertilisers. But, more recently, a rapid increase in the price of oil has, to some extent, hindered greater input of fertilisers. By 2000 A.D. the total requirement of fertilisers is expected to be around 14 to 16 million tonnes. This implies an incsease in fertiliser consumption by a factor of 3 over the present.

But the important point which we wish to emphasise is that merely an increase in fertilisers or an increase in irrigation and farming techniques would not make Indian agriculture entirely independent of rainfall in the coming decades. It looks more than likely that we will have to live with the dependence on rainfall for sometime. The crucial factor here seems to be our ability to anticipate likely changes in rainfall pattern from one year to another. This is essentially a scientific problem. Currently, with the techniques which we have described earlier, a forecast by statistical methods can be provided on a monthly basis for about 35 meteorological sub-divisions within the country. The percentage of success by these methods for a monthly forecast is around 65%. But it must be pointed out that each sub-division is divided into a number of districts. Thus, even if the rainfall for the subdivision as a whole might be either normal or in excess, there could be districts within the suh-division where the population is faced with inadequate rainfall. Apart from the distribution of rainfall, its timeliness is important. Thus, it may occur that although the cumulative rainfall was adequate its timeliness was bad. Prediction techniques that are currently available cannot forecast the time of occurrence of rainfall more than a few days ahead. The position is, unfortunately, still less satisfactory for longer forecasts

for a month ahead. Thus, with techniques which are available currently one might forecast with some success, which is better than what would be achieved by mere chance, about the cumulative rainfall for the month as a whole. But, this rainfall might occur in the last week of the month when it is least needed, while the first week of the month when rainfall was most needed, might be dry.

To get over this difficult problem, there is greater need for research on the dynamics of the monsoon. Considering the large expense of research on long range prediction one is tempted to visualise the formation of a central agency by pooling the resources of several individual organisations. Such a centre would need to be provided with a large computer with fast access time. One would need to construct a model monsoon and experiment with different facets of the model to identify those features which have the largest control over rainfall. As described in an earlier chapter, a beginning has been made in India but, clearly. much remains to be done. In particular, one must attempt greater integration between meteorological variables and crop patterns.

This aspect was considered in some detail by the National Commission for Agriculture in India in 1976. The Commission divided the country into five rainfall regimes based on the total quantum of likely rainfall in a three-month period. Their broad conclusions are shown in table 14.1.

Table 14.1
Rainfall classification and main crops

Category	*Rainfall amounts (in cm) per month for at least 3 consecutive months*	*Suitable crops*
A	More than 30	Paddy
B	20—30	Maize, Blackgram
C	10—20	Bajra, Millets
D	5—10	Grass (achitofolius), Fodder
E	Less than 5	Not Suitable for crop production

From this table, we note the rainfall of the same type is needed for at least a period of three consecutive months for the satisfactory growth of different crops. Unfortunately, it is only in

20% of the country that rainfall of the same type of distribution can be assured, with reasonable expectancy, for three consecutive monsoon months. Over the remaining 80% of the total area under crops, the rainfall distribution only partially satisfies crop requirements. Support from irrigation is at present limited to another 20-25% of the area under agriculture.

In broad terms, therefore, we need to concentrate attention on rainfall forecasts for roughly 50% or half the total land under cultivation. This has to be considered along with the fact that even during the monsoon season the variability of rainfall is high in the months of June and September.

The crops which are most suited to monsoon rainfall are millets. Paddy, which depends on large amounts of rainfall, suffers from the danger of water logging and floods unless suitable preventive measures are adopted at the very beginning. The time of sowing needs to be adjusted to the date of commencement of rainfall. This can ensure a reasonable supply of moisture to the soil at least in the early stages of the crop. The National Commission on Agriculture notes that "as a second line of defence one has to keep ready" a schedule of other crops which could be sown depending on the early or late occurrence of rainfall, so that the chance of a complete failure inherent in placing reliance only on one crop is minimised.

The National Commission also noted defects which occur through lack of communication between the farmer and weather information on rainfall. It has been mentioned, for example, that paddy is sometimes grown in regions where rainfall is highly variable. This leads to low and uncertain yields. A change over from paddy to crops like maize in such areas could increase the production by as much as 2-3 tonnes per hectare. There is considerable scope for diversifying the cropping pattern by introducing crops both before and after paddy in such regions. In low rainfall areas, it is desirable, the Commission point out, to grow a variety of crops without any single dominating crop. Jowar, bajra, ragi, groundnuts and grams are examples of cereals that may be grown in these regions.

In our country crop weather calendars have been designed by the Meteorological Department. These calendars indicate the

meteorological requirements at different stages of a plant's growth. Unfortunately, the greatest emphasis has been placed on rainfall. Apart from rainfall there are other meteorological elements, such as sunshine, temperature, humidity and soil moisture which have an impact on crop yields. There appears to be need for further experiments under controlled conditions in this area.

Green houses have been utilized to simulate different types of weather to study their impact on plant growth. It is possible that there might be an optimum combination of weather elements which could maximise crop production. Such experiments could provide information that would be useful for planning the diversion of water from irrigation channels, or for tapping ground water reserves in the event of inadequate rainfall.

Evapotranspiration is an another important factor for crop growth. Currently, theoretical models are used to assess evapotranspiration, but it is not yet very clear how far there models represent reality. Further experimentation is desirable.

Many agricultural pests and diseases are influenced by weather. It has been estimated that over 10% depletion of the crop could result from pests and diseases. This is another area where the study of weather and agriculture could pay rich dividends, especially on the movement of locusts.

It is of some interest to enquire into the type of questions to which a meteorologist can readily provide answers that will help the farmer. In general terms, we pose the following questions :

(*i*) Is the use of the land, or the result of an agricultural operation, dependent on weather ?

(*ii*) If the answer to (i) is in the affirmative, one would wish to know (a) the major weather events and (b) the biological stages in the growth of a plant when weather plays a crucial role.

(*iii*) How far in advance can meteorological events be predicted ?

(*iv*) What is the best way to exploit a weather event of known probability, such as the Indian monsoon ?

(*v*) Can we express the advantages of a correct weather prediction in financial terms ?

If we are to provide answers to these questions, we shall need more intensive research on the monsoon. The last question, and its answer, is of special interest to our economy. A few figures

that have been recently published by a panel of experts invited by the World Meteorological Organization are indicative of the type of results that could be achieved.

In the United Kingdom hay is a weather sensitive crop. The total annual crop is worth about £ 70 million. It has been estimated that the judicious use of weather forecasts in selecting the best times for cutting and drying the crop could increase the yield and quality of hay by 10%. This represents a benifit of £ 7 million.

The quality and quantity of hay are important to milk praduction, which is worth about £ 200 million. A poor hay crop can reduce this by about 2 per cent, implying a loss of about £ 4 million. Thus, the meteorological service for milk and hay production is worth about £ 11 million.

In a similar manner, the net gain can be computed, in a fairly realistic manner, for other services to agriculture in the United Kingdom. The incidence of liver fluke in sheep, for example, can be forecast with the help of meteorological data. If suitable veterinary action is taken, the net gain is of the order of £ 1 million. Meteorological forecasts based on past weather can indicate the outbreak of virus diseases in sugar beet. This produces a benefit of another one million pounds per year. If we add similar benefits, the total benefit to agriculture may be placed somewhere between ten and twenty million pounds per year.

It has been estimated that the staff and organization provided for agricultural meteorology costs the British Exchequer between £ 50,000—80,000 per year. On the basis of these figures, the benefit cost ratio is :

(10 million to 20 million)—(50,000 to 80,000).

This suggests a benefit/cost ratio of at least 100 : 1, and the figure may well be twice as large. In India, which is primarily an agricultural country, it would not be unrealistic to assume similar results.

Another active field in which the monsoon plays a vital role is the proper utilization of our water resources. Water is one of India's valuable natural resources but, if not properly controlled or used wastefully, it can cause considerable damage. Meteorological studies of the monsoon can help in making the best use of our water resources. Rainfall and run-off data are necessary to

determine when the flow of water in rivers must be restricted; measures for flood control or warnings for uncontrolled floods need prediction of heavy rainfall.

It is estimated that in the United States the annual savings from flood warnings have averaged more than $ 30 million per annum. In India, if we are able to save even a tenth of this figure, it implies a saving of about Rs. 2 crores per year.

In this concluding chapter we have presented an account of a few activities in which considerable benefits may be derived from meteorological advice. The ultimate aim of this survey is not to present figures that are flattering to a meteorologist; on the contrary, it is our intention to stress the need for better utilization of our natural resources by judicious use of weather information. A recent study arrived at the conclusion that the annual potential benefits to the world from World Weather Watch would exceed £ 16 billion. This is about 50 times the estimated yearly cost of implementing this project. It is by no means too early for us to start thinking of the opportunities that lie ahead.

REFERENCES

CHAPTER I

1.	*Blanford, H.F.*	Climates and weather of India, Ceylon and Burma, Macmillan Company, New York, 1889.	10
2.	*Chatterji, G.*	An improved method of sounding the lower layers of the atmosphere, Ind. Met. Dept. Sci. Notes, 35, 4 (1931).	12
3.	*Dines, W.H.*	Air Ministry, Meteorological Office, M.O. 321, London, (1929).	11
4.	*Hadley, G.*	Concerning the cause of the general trade-winds, Phil. Trans. Royal Soc., London (A), 29, 58-62 (1735).	4
5.	*Halley, E.*	An historical account of the trade-winds and monsoons observable in the seas between and near the tropics with an attempt to assign the physical cause of said winds, Phil. Trans. Royal Soc. London 26, 153-168 (1686).	4
6.	*Mahalanobis, P. C.*	Rain storms and river floods in Orissa, Sankhya, Calcutta, 5, 1, 120	12
7.	*Ramage, C.S.*	Monsoon Meteorology, Acad. Press, New York, 296 pp (1971).	4
8.	*Unakar, M.V.*	Correlation between pre-monsoon Conditions over N.W. India and subsequent monsoon rainfall over N.W. India and the Peninsula.	1

Ind. Met. Dept. Sci. Note Vol. I, No. 6 (1928).

9. *Walker, G.T.* On the meteorological evidence for supposed changes of climate in India Mem. Ind. Met. Dept. 21, 1, 22 pp. (1910). 10

CHAPTER II

10. *Bjerknes, J.* Atmospheric teleconnections from the equatorial Pacific, Mon. Wea. Rev., 97, 163-172 (1969). 30

11. *Budyko, M.I.* Atlas of the heat balance of the Globe, Hydrometeorological Service, Moscow, USSR, 69 pp (1963) 23

12. *Webster, P.J.* A model of seasonally—varying planetary scale monsoons, *Monsoon Dynamics*. Ed. Sir James Lighthill and Prof. R.P. Pearce, Camb. Univ. Press. 165-191 (1981). 22

CHAPTER III

13. *Doraiswamy Iyer, V.* Typhoons and Indian Weather, Mem. Ind. Meteor, Dept., 26, Part VI., 93-130 (1935). 34

14. *Findlater, J.* An experiment in monitring cross-equatorial airflow at low levels over Kenya and ratnfall of western India during the northern summer, *Monsoon Dynamics*, Ed. Sir James Lighthill and R.P. Pearce, Camb. Univ. Press, 309-318 (1981). 39

15. *Lighthill. M.J.* Dynamic response of the Indian Ocean to the onset of the southwest monsoon Phil. Trans. Roy, Soc, London A-265, 45 pp (1969). 42

16. *Rossby, C.G. et al.* Relation between variations in the intensity of the zonal circulation of the atmosphere and the displace- 44

ments of the semi-permanent centres of action, J. Mar. Res., 2, 38-55 (1939).

17. *Saha, K.R., J. Shukla and F. Sanders* Westward propagating predecessors of monsoon depressions, Mon. Wea. Rev., U.S.A. 139, 2, 330-343 (1981). 35

18. *Simpson, G.C.* The southwest moosoon, Quart. J. Roy. Met. Soc., London, Vol. XL VII, 199, 151-172 (1921). 44

CHAPTER IV

19. *Koteswaram, P.* An analysis of the high tropospheric wind circulation over India in winter, Ind. J. Met. Geophys., 4, 13 pp (1953). 62

20. *Krishna Rao. P.R.* Some high sounding ballon ascents and upper air temperatures upto 35 km. over India, Ind. J. Met. Geophys., 3. 173 pp (1952). 62

21. *Rossby, C.G.* On the maintenance of the westerlies south of the polar front. Fluid mechanics applied to the study of atmospheric circulation, Pap. Phys. Oceonog. Meteor., U.S.A. 7, 1, 9-17 (1938). 62

CHAPTER V

22. *Krishna Rao, P.R. and P. Jagannathan* A study of the N.E. monsoon rainfall of Tamilnadu, Ind. J. Met. Geophys., 4, 22 pp (1953). 71

CHAPTER VI

23. *Bannerji, S.K. and A.K. Mukherji* On methods of exciting nature's process for production of rain, Proc. Symp. on Artificial Rain, CSIR, New Delhi, 40-77 (1955). 90

24. *Bergeron, T.* On the physics of cloud and precipitation, Proc, 5th assembly, UGGI, 82

25. *Findeisen W.*	Entstehn die Kondensationskerne an der Meeresoderflache, Met, Zeit, 54, 377-379 (1937).	82
26.. *Kelvin, Lord*	On the equilibrium of vapour at a curved surface of a liquid, Proc. R. Soc. Edinburough., 7, 63 (1870).	78
27. *Kraus, E.B. and P. Squires*	Experiments on the stimulation of clouds to produce rain, Nature, London. 159. 489 (1947).	90
28. *Mason B.J.*	*The physics of clouds*, Clarendon Press, Oxford, (1957).	81
29. *Ramana Murty, Bh. V. at al.*	Results of cloud seeding experiments at Delhi as assessed by Radar, Ind. J. Met. Geophys., 20, 11, (1969).	90
30. *Roy, A.K. et al.*	Cloud seeding trials at Delhi during monsoon months, July to September (1957-59), Ind. J. Met. Geophys., 12, 401 (1961)	90
31. *Schaefer,V.J.*	The production of ice crystals in a cloud of supercooled water droplets. Sci., New York, 104, 457 (1946).	89
32. *Schaefer,V.J. and I.Langmuir*	First quarterly pregress report, Meteorological Research, General Electric Res. Lab., U.S.A. 15 July, 1947.	89
33. *Simpson, G.C.*	On the formation of clouds and rain, Quart. J. Roy. Met. Soc., London, 67, 99-133 (1941).	81
34. *Wilson, C.T.R.*	Condensation of water vapour in the presence of dustfree air and other gases, Phil. Trans. Roy., London A 189, 565 (1897).	79

CHAPTER VIII

35. *Bjerknes, V.*	Dynamic Meteorology and Hydrography, Chapter. VI, Carnegie Institution of Washington, USA (1910).	110

36. *Charney, J.G.* On a physical basis for numerical prediction of large-scale motions in atmosphere, J. Meteor., U.S.A. 6, 371-385 (1949). 108

37. *Das P.K. and H.S. Bedi* Proc. Ind. Acad. Sci. (Earth & Planet Sci), 89, 2, 159-177 (1980). 112

38. *Obukhov, A.M.* The problem of geostrophic winds, Izy, Akad. Nauk, USSR, Ser. Geogr. i Geofyz., Vol. 13,4, 281-306 (1949). 110

39. *Ramage, C.S.* *Monsoon Meteorology*, Acad. Press, New York, 296 pp (1971). 119

40. *Richardson, L.F.* Weather prediction by numerical process, Camb. Univ. Press, 236 pp (1922). 110

41. *Rossby, C.G.* The scientific basis of modern meteorology, Climate and Man, Yearbook of Agriculture, Washington, U.S. Govt. Printing Off., 599-655 (1941). 110

42. *Shukla, J.* Effects of Arabian sea-surface temperature anomaly on Indian summer monsoon : A numerical experiment with the GFDL Model, J. Atmos. Sci., 32, U.S.A. 503-511 (1975). 116

CHAPTER IX

43. *Bryson, R.A. and D.A. Barreis* Possibilities of major climatic modification and their implication over Northwest India, a case for study, Bull. Amer. Met. Soc., 48 (3), 136-142 (1967). 122

44. *Charney, J.G.* Dynamics of deserts and drought in the Sahel, Quart. J. Roy. Met. Sec., London 101, 193-202 (1975). 126

45. *Koppen, W.* Grundriss der klimakund, Berlin : Walter de Gruyter Company (1931). 121

46. *Singh, G.* Birbal Sahni Institute of Palaeobotany, Lucknow (Persanal communication). 122

CHAPTER X

47. *Bunker, A.F.* Interaction of summer monsoon air with the Arabian sea (Invited lecture), Proc. Symp. on Int. Ind. Ocean Expd., Bombay, 3-16 (1965) 137

48. *Ferrel, W.* Nashville J. Med. and Surgery, Prof. Pap., Signal Serv., USA, 12, 7, (1856). 143

49. *Ghosh, S.K. et al.* Influence of the Arabian sea on the Indian summer monsoon, Tellus, 30, 117-125 (1978). 139

50. *Houghton, H.G.* On the annual heat balance of the Northern Hemisphere, J. Met. USA, 11 (1) : 1-9 (1954). 136

51. *Lorenz, E.N.* The nature and theory of the general circulation of the atmosphere, WMO, Geneva, 1-1161 (1967). 143

52. *Pisharoty, P.R.* Evaporation from the Arabian sea and the Indian southwest monsoon, Proc. Symp. on IIOE, Bomboy, 43-54 (1965). 138

53. *Rossby, C.G.* On the distribution of angular velocity in gaseous envelopes under the influence of large-scale horizontal mixing processes, Bull. Amer. Met, Soc., 21, 53-68 (1947). 146

54. *Simpson, C.G.* Further studies in terrestrial radi ation, Mem. Roy. Met. Soc., London, 3, 21, 1-26 (1928). 136

CHAPTER XI

55. *Sarkar, R.P.* Some modifications in a dynamical model of orographic rainfall, Mon. 159

Wea. Rev., Washington, 95, 10, 673-684 (1967).

CHAPTER XII

56. *Walker, G.T.* Correlation in seasonal variation of weather, VIII ; IX, Mem. Ind. Met. Dept., 24, 275-332 (1924). 162

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57. *Linsley, R.K., Jr. M.A.Kohler and J.L.H. Paulhus* *Hydrology for Engineers*, McGraw-Hill Book Company Second Edition (1975). 187

58. *McAlpin, M.P.* Death, Famine Risk : The changing impact of crop failures in western India, 1870-1920, J. Economic History, XXXIX, Tufts Univ., USA, 143-167 (1979). 191

59. *Sarkar, R.P.* Droughts in India and their predictability, Proc. Symp. on Hyd. Aspects of Droughts, IIT, Delhi, 33-40 (1979). 189

CHAPTER XIV

60. *Bhatia, B.M.* Famines in India, Second Edition, Asia Publishing House, Delhi, 389 pp (1967). 195

GLOSSARY

Angstrom : Unit of length used to measure the wavelength of solar radiation. 1 Angstrom$=10^{-8}$ cm.

Angular Momentum : Product of the momentum of a rotating body and its distance from the axis of rotation.

Anticyclone : A centre of high barometric pressure. The circulation round an anticyclone is clockwise in the Northern Hemisphere, and anticlockwise in the Southern Hemisphere.

Buys Ballot's law : A law relating the flow of wind with the pressure. According to this law, an observer standing with his back to the wind in the Northern Hemisphere will have higher pressure to his right and lower pressure to his left. In the Southern Hemisphere the opposite situation prevails.

Coriolis Force : A fictitious force caused by the rotation of the earth. It tends to deflect a moving body to the right of its path in the Northern Hemisphere.

Cyclone : A centre of low barometric pressure. The flow round a cyclone is anticlockwise in the Northern Hemisphere and clockwise in the Southern Hemisphere.

Correlation Coefficient : A measure of the association between two variables.

Intertropical front : An imaginary line of demarcation between the trade winds of the Northern and Southern Hemispheres.

Jet Stream : A narrow stream of very strong winds, usually observed in the high troposphere.

Kinetic Energy : Energy of motion measured by half the product of the mass of a moving body and the square of its velocity.

Lapse Rate : Vertical gradient of temperature in the atmosphere.

Micron : A unit used to measure very small lengths. It is a millionth part of a metre.

Momentum : Product of the mass of a body in motion and its velocity.

Nuclei : Small aerosols floating in the atmosphere on which condensation of water vapour takes place.

Potential Energy : Energy of form. It is measured by the product of *g*, the acceleration due to gravity, and the height of the body above sea level.

Stratosphere : A region of the atmosphere where the temperature remains constant or increases slightly with height.

Troposphere : The lower regions of the atmosphere where the temperature decreases with height.

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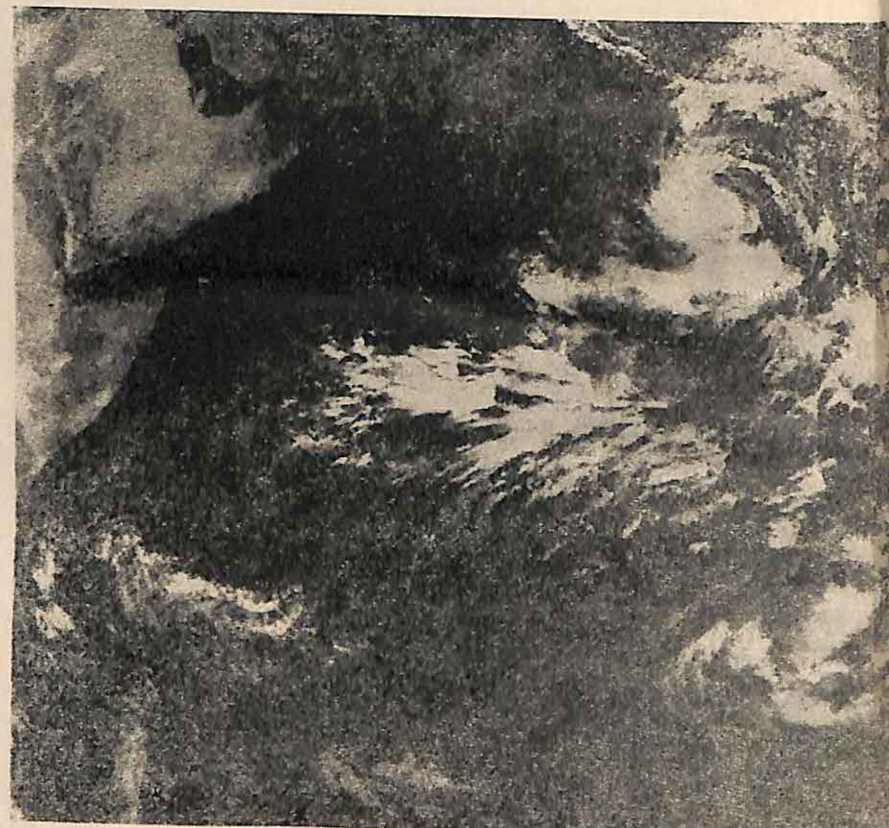

The reader will find here an account of how the genesis of the monsoons was traced with constant level balloons during the Monsoon Experiment of 1979 : how weather satellites are used to locate cyclonic storms over the Indian ocean, and, finally, how the climate of the Rajasthan desert may be made more friendly. This book should be of interest to all who would like to know more about the Monsoon, which is so much a part of our national life and heritage.

Dr. P. K. Das was the Director General of the National Meteorological Service of the Government of India. He is the author of several research publications concerned with the Indian Monsoon. He is a recipient of the third I.J.M.G. Award, a prize given biennially to the author of the best paper published in the Indian Journal of Meteorology and Geophysics (now named Mausam). Dr. Das was closely associated with many international scientific societies and organisations dealing with Meteorology. He was the first Indian to be invited to deliver the prestigious IMO Lecture on Monsoons by the World Meteorological Organisation (WMO) at its Fifth Congress in Geneva in 1983.

After retiring from the Meteorological Department in India, Dr. Das was Senior UNDP Expert in the Department of Meteorology at the University of Nairobi in Kenya. He is currently Professor of Meteorology at the Indian Institute of Technology in New Delhi.

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